

ICES Marine Science Symposia

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Volume 195 December 1992

Hydrobiological Variability in the ICES Area, 1980–1989

A Symposium held in Mariehamn, 5–7 June 1991

International Council for the Exploration of the Sea
Conseil International pour l'Exploration de la Mer

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Papers to celebrate the 70th Anniversaries of Stig Fonselius, Arthur Lee, Lars Midttun, Odd Sælen, and Warren Wooster

Edited by R. R. Dickson, P. Mälkki, G. Radach, R. Sætre, and M. P. Sissenwine

International Council for the Exploration of the Sea
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HYDROBIOLOGICAL VARIABILITY IN THE ICES AREA, 1980–1989

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Hydrobiological Variability in the ICES Area, 1980–1989

Papers to celebrate the 70th Anniversaries of Stig Fonselius, Arthur Lee, Lars Midttun, Odd Sælen, and Warren Wooster

Introduction

In October 1987, during the 75th Statutory Meeting in Santander, the Council endorsed the conclusion of its Consultative Committee that annual publication of *Annales Biologiques* was no longer feasible within the available budget, and Volume 41 for the year 1984 became the last in this distinguished series.

From the expressions of regret recorded by many Standing Committees at the discontinuation of this series (*Procès-Verbal*, 1987), it is plain that the *Annales* had been providing a compilation of information that was of some considerable value and not readily available elsewhere, and in such a situation, where the scientific justification remains but the costs of compilation and publication prove prohibitive, there was thought to be great point in seeking to meet the same scientific purpose via a different and more affordable route.

On the advice of a small working party, consisting of the Chairmen of the Hydrography, Marine Environmental Quality, and Biological Oceanography Committees, together with the Hydrographer and Environment Officer, and taking note also of the discussions of this question within the main Fish Committees, the Consultative Committee recommended that the essential purpose of *Annales Biologiques* might be met by promoting three quite different activities within ICES:

(a) actively solicit interpretative papers on time series of data on physical, chemical and biological environmental parameters for publication in the IOC Time Series on Ocean Measurements;
(b) promote symposia or theme sessions on "Overviews of the Decade" following the approach of NAFO. The first of these decadal overviews, which could be published by ICES in the Cooperative Research Report series, should be held in 1990 or 1991 to cover the decade of the 1980s;
(c) give a high priority to the transfer of environmental data from member countries to the ICES permanent data bank and to quality control exercises that enhance data comparability in time and space.

This recommendation was accepted by the Council, and at the 76th Statutory Meeting in Bergen the Council gave it substance by recommending that a Symposium on "*Hydrobiological Variability in the ICES Area, 1980–1989*" should be held in Mariehamn in the Baltic during June 1991 (C. Res. 1988/2.2). The present volume is the scientific record of these proceedings.

While the above describes the bare facts of the decision, it describes little else. Yet, in the case of this particular subject, where the issue is the recognition and interpretation of long-term change, the detailed context of these decisions is of a peculiar importance. It matters, in other words, why the Symposium should have been held in Mariehamn and in 1991; it matters what the original aims of *Annales Biologiques* were, and who drafted them; and it matters too that the Symposium's participants should intend these proceedings to form a Festschrift to commemorate the 70th anniversaries of five of their colleagues.

In each case the reason is one of *continuity*. It is a sobering fact, captured by Duarte *et al.* (1992) that although there has been an exponential increase in the initiation of new time-series-monitoring programmes in European marine stations in recent decades, there has also been an exponential increase in their termination, so that "long-term monitoring programmes are, paradoxically, among the shortest projects in marine sciences: many are initiated, but few survive a decade". Put differently, policy-makers are rather easily startled into initiating actions, but the time scales of policy, funding, and even of scientific "fashion" are not normally imbued with much stamina.

"Consequently", as Duarte *et al.* point out, "the continuation of long-term monitoring programmes is often heavily dependent on the personal effort and dedication of individual scientists", and nowhere is the truth of this statement better recognized than in the Standing Committees of ICES, where individuals and groups of individuals have long taken the responsibility for piecing together the evidence for change in their regional marine environments and for attributing these changes as to cause.

If we sometimes appear to be unduly preoccupied with the personalities and history of this subject within ICES, it is a preoccupation that is readily explained – as

one practical way of acknowledging and encouraging these efforts.

Seen in this light, the Mariehamn Symposium is the end-point of actions by three main groups of individuals over a period of more than a century. First, the location of the Symposium on an island in the Baltic and its timing are a deliberate reference to the fact that one hundred years earlier, in 1890, Otto Pettersson and Gustav Ekman began the process of hydrographic monitoring in the Baltic and North Seas, working agreed routes and stations on a quarterly basis. In 1892, this was developed into an international scheme of investigations at the request of a meeting of Scandinavian naturalists held in Copenhagen, and in 1893 this international effort got under way, with scientists from Sweden, Norway, Denmark, Germany, and Scotland taking part. From the outset, these studies had a multidisciplinary, "applied" aim, with plankton observations being included and with Pettersson attempting to show the influence on the Bohuslän herring fishery of changes in the hydrography. The utility of Pettersson's scheme prompted recommendations that the Swedish Government should initiate broader intergovernmental actions based on it, and following the International Conferences on Fishery and Hydrographical Investigations in the North Sea, held at Stockholm and Kristiania in 1899 and 1901, the International Council for the Exploration of the Sea was founded in 1902.

A second, later set of actions acknowledged in the design of the Mariehamn Symposium are those which led to the founding of *Annales Biologiques* in 1939–1943. It could hardly be otherwise. As the decadal conference designed to replace the *Annales*, the content of this Symposium had to be designed to meet the essence of the original objectives. In all probability, the initiation of the new *Annales* in 1943 at such an inauspicious time in world affairs must reflect the unprecedented nature of the hydrobiological changes which were forcing themselves on to the scientific consciousness at that time, not merely the energy and foresightedness of the scientists involved. For these were the decades of "the warming in the north", the years when the salinity of the North Atlantic Water passing through the Faroe–Shetland Channel reached a century-long high (Dooley *et al.*, 1984), and the years when the northward dislocations of biogeographical boundaries for a wide range of species from plankton to commercial fish (not to mention terrestrial mammals and birds) were at their most extreme in the present century. The astonishing nature of these radical events is vivid in the contemporary scientific literature, most notably in the classic accounts by A. S. Jensen (1939), Å. Vedel Tåning (1943, 1949), B. Sæmundsson (1934), A. Fridriksson (1949), N. M. Knipowitsch (1931), A. C. Stephen (1938), and many others, summarized in a comprehensive bibliography by Arthur Lee (1948).

A swordfish off Iceland! Tunny in small shoals off the Icelandic south and west coasts (Fridriksson, 1949). Salinities so high off Cape Farewell that in 1933 they were thrown out as erroneous (Harvey, 1962). Precipitous warming by more than 2°C in the five-year mean all along the West Greenland Banks. And, apparently in response, the beginnings of an inexorable rise in the West Greenland cod catch from small beginnings in the 1920s through 300 000 t/a in the 1950s, and ultimately to more than 450 000 t/a attained in the 1960s before renewed cooling reversed the process (Buch and Hansen, 1988).

With these riveting events preoccupying the foremost figures of the day, it is little wonder that their enthusiasm would be transmitted to a whole new generation of their younger staff, together with the realization that the physical environment was subject to radical change over time scales of years and decades. Today, this point would be regarded as self-evident, yet it was not always so universally accepted. Compare today's *expectations* of global change and assertion of global warming – even prior to its demonstration – with the caution evident in Hans W. Ahlmann's remarks when he presented the Introductory Address to the ICES Special Meeting on "Climatic Changes in the Arctic in Relation to Plants and Animals" in 1948: "Prof Keranen says it is no longer possible to deny that the climate is improving, even though this is contrary to the thesis of the old meteorological school concerning the stability of the present climate." And it was not much earlier that Kincer (1933) could still ask "Is our climate changing?"

Such old-fashioned notions could hardly survive long in the face of such obvious and radical changes in both the circulation and ecosystem of the subpolar gyre. And it did not take long for the energetic and capable individuals of the day to respond. In his Preface to the first volume of the new *Annales Biologiques* series, the then ICES President Johan Hjort manages beautifully to capture this watershed in scientific thought and the progression in our thinking which had necessitated the new series when he refers to "the hypothesis, which at the beginning had seemed so daring . . . that the statistics of the catches of the fishermen or research vessels may be considered as representative of the existing stock"; that "samples collected from a series of years disclosed the important fact of fluctuations in the stock from one year to another"; and that "from the closer understanding of these changes arose the conviction that such changes were governed . . . by the influence of the environment which might be investigated not only for a rational understanding of the contemporary situation but also for prognoses as to events in future".

Though the idea of a permanent International Commission to make and administer such prognoses was proposed in the Berlin Meeting of the International Council, such a degree of official international cooperation proved impossible in the climate of May 1939. At the level of the individual scientist, however, there

was no such problem, and the Consultative Committee under Chairman E. S. Russell, Vice Chairmen A. Hagmeier and E. Le Danois, and its Sub-Committee on "Hydrographical and Biological Investigations" under the Chairmanship of M. Knudsen went right ahead with a plan to collect the biological and environmental data that such an International Commission *would* one day require when hostilities ceased.

Annales Biologiques was the expression of that plan; and with an energetic Jens Smed returning from his Master's degree in physics in 1939 to play his key dual role in developing the Service Hydrographique as an effective regional data centre, and in piecing together the accumulated data into reliable long time series, the plan was rapidly equipped with the necessary physical infrastructure to match the data products of the Fish Committees.

These actions, however necessary and effective, still do not add up to the long time series of multidisciplinary data against which the slow shifts of environmental change can be recognized and explained. To achieve *that* required simple continuity of enthusiasm and effort while the time series and data sets lengthened.

And so the third group of actions that the Mariehamn Symposium was designed to acknowledge were the actions of the individuals who had taken part in the lengthening and broadening of our description of the marine environment throughout the post-war years, and who are still fired with the idea that motivated the *Annales* in the first place – namely, that changes in the ecosystem and the physical environment were linked in some understandable way.

In many ways their task was the hardest of all, as the certainty with which Hjort and his contemporaries viewed the connection between the great fisheries and their physical environment sometimes proved difficult to pin down, as the time scale of Government policies and the stamina of their funding both shortened, and even to some extent as the fashions of science switched away from the long haul of monitoring and description to the short-term intensive study of process.

The five individuals whose efforts were especially acknowledged in the Mariehamn Symposium, and whose 70th birthdays are celebrated in this Festschrift, were both leaders and long-time contributors to the study of environmental variation and effects on biota over many decades within ICES. It is perhaps one of the best design-features of the "decadal symposium" idea that it will offer us a recurring opportunity to acknowledge and encourage similar efforts by others into the future.

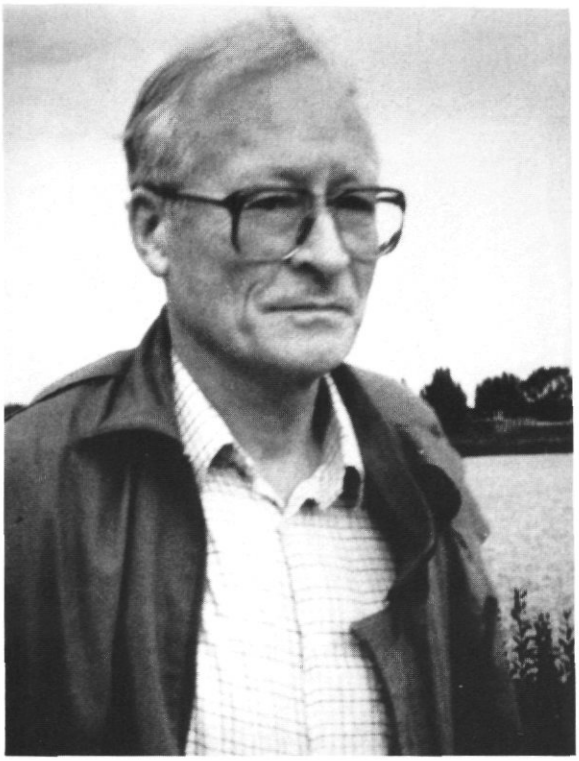
For the present these individuals will forgive us if, with five long careers to consider, we are forced to describe them in only a few inadequate sentences. We do so in order of age rather than alphabet.

Continued overleaf



Arthur Lee

Arthur James Lee (born 17 May 1920) joined the Hydrographic section of the Fisheries Laboratory, Lowestoft, from Cambridge University in 1947 and became a part of what was known as the Arctic Team, with a brief to study the hydrographic and climatic environment of the Barents Sea in relation to local fisheries, and where, most notably, he was responsible for the intensive working of the west Bear Island and Bear Island–Fugløy standard sections at quarterly intervals for a period of more than a decade from 1949. In the manuscript of a paper written for his Special Lecture to the 1970 Statutory Meeting (Lee, unpublished) he recalls the importance of his first scientific meeting – the 1948 ICES Special Meeting on Climatic Changes in the Arctic in Relation to Plants and Animals – in sparking a life-long interest in environment:fishery relations, and in the time series which describe the scales and effects of environmental variation. On his route to becoming Director of Fisheries Research from 1974 to 1980, these interests found expression in his role as International Coordinator for the NORWESTLANT Surveys of ICNAF in 1965; in his involvement in the pilot experiment for the international communication of oceanographic observations in 1966; in his role as co-founder in 1970 of the Joint North Sea Information System (JONSIS) which had the aim of establishing arrays of moored long-term oceanographic data stations in the North Sea, and under which the first quasi-continuous stations of a decade's duration – JONSIS 1 and 2 – were actually established by the Lowestoft Laboratory off the Yorkshire Coast; in his support for the two offshoot Data Acquisition programmes, JONSDAP '73 and '76, one of whose phases finally implemented a plan developed under Hjort in 1940 to measure the variability of the flux of Atlantic Water into the North Sea and Norwegian Sea; in his administrative role in ICES as Chairman, Hydrography Committee (1968–1970), Chairman, Consultative Committee and Liaison Committee (now ACFM) (1970–1974), and Chairman, Advisory Committee on Marine Pollution (1972–1974); and throughout this period, his continuation of the same interest in the fluctuations in marine climate and fisheries with which he started his publication list as compiler of the exhaustive bibliography for the 1948 Special Meeting, and to which he was still contributing in 1988, while actively engaged in piecing together the evidence by which the "Great Salinity Anomaly" came to be understood. It is a nice point that the cool, fresh conditions of that record event essentially ended the great period of "the warming in the North" with which his career had begun.



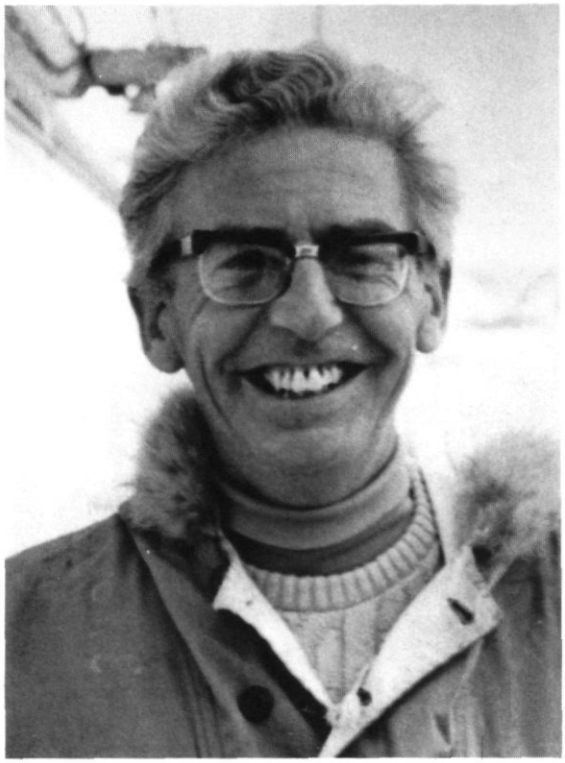
Odd Sælen

Following his degree in Oceanography at the University of Oslo in December 1945, **Odd Hendrik Sælen** (born 8 June 1920) describes his first position as that of "amanuensis" to Professor Håkon Mosby at the Geophysical Institute in Bergen in 1947.

However, to the rest of us, these were the years in which he began his classic studies of hydrographic variability on the standard Sognefjord Section, published in 1959 and 1963 and for which he earned his Doctorate. These time series of transport calculations for the period 1947–1959, which included some direct current measurements using Mosby's bottom current meter in 1957 and 1959, were the direct descendants of earlier workings of the Sognefjord Section by Helland-Hansen and Nansen in 1901–1905 and by Helland-Hansen in 1925–1929. However, the intensity of working this section (27 times between 1947 and 1953, for example) was quite new, so that the high-frequency variability in the Norwegian Atlantic Current was for the first time adequately described and set into the context of longer period fluctuations; the "puzzling waves" in the isosteres which had baffled Helland-Hansen and Nansen were shown, with the use of multi-ship surveys, to be the two-dimensional expressions of intense vortices in the Atlantic current; and we began to see both the effects on currents of strong meteorological disturbances but also the effect on variability of vertical movements of cold water up the Slope. In short, our thinking became modernized.

Following a period as Professor in succession to Fjeldstad at the University of Oslo from 1965, he returned as Professor of Oceanography to the University of Bergen in 1978. He served as Chairman of the ICES Hydrography Committee (1970–1973), Chairman of the Norwegian Oceanographic Committee (1971–1974), and Chairman of the Nordic Council for Physical Oceanography (1978–1981).

While his studies of time series also extended to the ventilation history of Norwegian fjords and while his bottom-current measurements were to continue with the first direct measurements of bottom-water exchange between the Greenland and Norwegian Seas, Odd Sælen would perhaps forgive us for suggesting that his greatest legacy – along with others who have demonstrated the utility of standard-section data since the time of Helland-Hansen and Nansen – lies in the continuation of the Norwegian Standard Sections programme, which still routinely provides the best characterization of physical variability in our sector of the North Atlantic and which has underpinned many of the papers presented at this Symposium.



Lars Midttun

Lars Sudmann Midttun (born 16 June 1920) began his scientific career as assistant to Professor Bjørn Helland-Hansen at the Geophysical Institute in Bergen in 1944–1945. In 1947–1948, after receiving a degree from the University of Oslo, he became a member of the Norwegian Antarctic Expedition aboard "Brategg". In 1948–1953, as a scientific assistant to Professor Mosby at the University of Bergen, he began the work on sound propagation in the sea which led to a long involvement with the field of fisheries acoustics.

However, his involvement with the subjects which form the basis of this Symposium – fisheries oceanography, long time series and impacts of climatic change – really began when he joined the Institute of Marine Research in Bergen (from 1953 as member of staff and since 1969 as Head of the Physical Oceanography Division). At that time, Rikard Ljøen was given responsibility for physical variability studies in the North Sea, Jens Eggvin, the Norwegian Sea, and Lars Midttun, the Barents Sea, and it has been his principal aim and contribution to maintain the quality, consistency, and continuity of the data set from these northern seas ever since.

From his 1969 paper, in which the hydrographic variability at a range of Norwegian sites is partitioned into its short-term, seasonal, and interannual components, we progress in understanding through his 1975 analysis of 35-year T and S series at a complete ring of 25 coastal sites (many begun by Eggvin) and ultimately to the point (Midttun *et al.*, 1981; Midttun and Loeng, 1987; Blindheim and Skjoldal, in press) where the geographical distribution of different age groups of commercial fish in the Barents Sea is linked to the fluctuating state of their environment. Finally, Hjort's goal of *prediction* seems in prospect; but in the most advanced model yet produced of the climatic system in the Barents Sea (Ådlandsvik and Loeng, 1991) it is no surprise to note that the inflow that they model may well be driven in part by the outflows of dense bottom water from the eastern Barents Sea on which Lars Midttun's most recent contributions have been focused.



Warren Wooster

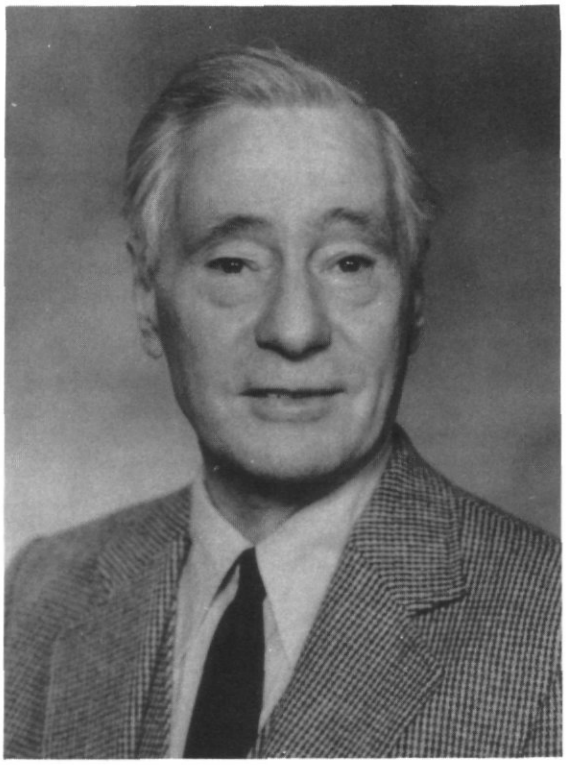
Warren S. Wooster (born 20 February 1921) initiated his career and formed his enduring enthusiasm for the study of interactions between the physical environment and the ecosystem as a young postgraduate chemist investigating "Phosphate in the eastern North Pacific" for the Californian Cooperative Fisheries Investigations (CalCOFI) programme.

During a long and distinguished academic career at the UCSD – Scripps Institution of Oceanography, University of Miami, and University of Washington, he found the time to serve as Secretary of the Intergovernmental Oceanographic Commission (1961–1963), President of SCOR (1968–1972), and President of ICES (1982–1985) and is even now the motive force behind the establishment of a Pacific version of ICES, to be known as PICES.

While such an illustrious series of posts in the administration of marine science could not be omitted from any pen-picture, however brief, his research interests are of even greater relevance, and here, by contrast, and to our great benefit, he has travelled very little away from the physical, chemical, and ecosystem exploration which made up the enthusiasms of his CalCOFI days.

His early work into the dynamics of upwelling systems off the west coasts of North and South America, with Joe Reid into the physics and chemistry of eastern boundary currents, with Townsend Cromwell and Gilmartin in describing a whole new system of undercurrent and counter-current in the equatorial Pacific during the 1960s, and with Schaeffer and Robinson in compiling their atlas of the Arabian Sea in 1967, were important not only in themselves, but were continually referenced against the dynamics of the local ecosystem. The atlas, for example, was the first thorough and consistent evidence for the great productivity of the Arabian Sea, while the East Pacific studies formed an important part of the physical context through which we can now understand the extreme coherence of variability in the planktonic ecosystem of the California Current, and the response of that ecosystem to imposed climatic signals as large as those of El Niño.

The key point is that in 1988, with collaborators in ICES, he was still describing the physical basis for planktonic variation and ecosystem change, this time in the eastern Atlantic and using the Continuous Plankton Recorder (CPR) data set, not that of CalCOFI; that today he is still contributing (even at a Symposium in his honour!) to the links between ecosystem change and fisheries in the eastern Pacific; and that when the very existence of the CPR survey was in question, it was the old CalCOFI chemist who had the interest to take responsibility for a US part in the rescue of the Survey and its venerable time series.



Stig Fonselius

The youngest of our five septuagenarians, **Stig Henrik Johannes Fonselius** (born 29 April 1921), progressed from a chemistry degree and M.Sc. at the University of Helsinki in 1945–1951 to work initially as an assistant in chemical meteorology at the University of Stockholm under the direction of Professor Carl Gustav Rossby (1954–1959), and from there to begin his long and varied career with the National Board of Fisheries, as hydrographical assistant, Inspector of Fisheries, head of the Chemistry Department, and, from 1979 till "retirement" in 1987, as Director of the Oceanographical Laboratory of SMHI.

Throughout this long period, his main research interests have been twofold. First has been his lifelong involvement with the study of stagnant conditions in the Baltic, which led very rapidly to the production of three classic works on "The Hydrography of the Baltic Deep Basins", I, II and III in 1962–1969 and for which he was awarded his Doctorate by the University of Göteborg in 1969. In this, our story comes full circle because, as he points out in the third of these studies, his chemical fieldwork was the continuation not only of earlier work by Kalle and others in the 1930s and 1940s but also of the earliest chemical (oxygen, CO_2 and nitrogen) analysis that O. Pettersson had conducted in 1898 as part of his first Baltic Monitoring Programme.

In his own contributions, however, Stig Fonselius was not merely adding longevity to an existing record but, because of the recent salinification of the Baltic Deeps, was describing a quite new situation in which effective ventilating inflows were necessarily of such a high density that they themselves might usher in stagnation for a decade or more. Thus, although the first stagnation described by Kalle lasted for four years (from 1928), the great inflow of 1951 caused a stagnation which was not broken until 1961 in the Gotland Deep, and, in the 1992 report of the comprehensive SMHI Program for National Open Sea Monitoring (Program för Miljökvalitet, or PMK), we learn that in the latest stagnation hydrogen sulphide has now been continuously present in the bottom waters of the Gotland Deep for more than 13 years.

The second of Stig Fonselius's research interests has been more closely allied to the third suggestion of the Consultative Committee for the replacement of *Annales Biologiques*, namely the establishment of "quality control exercises that enhance data comparability in time and space". In his long series of reports on the standardization of methods for oxygen determination (June 1965, September 1966), phosphate (May 1966), chemical methods generally (FAO, 1975), and in his close involvement with the intercomparison of operational hydrographic measurements during the multi-ship surveys of BOSEX '77, PEX '86, and SKAGEX '90 and '91, it is plain that he has already anticipated that requirement.

Despite the inadequacy of these few remarks to describe such very different people, it is hoped at least that the common thread shows through – namely, that each has been concerned with the increasingly difficult task of maintaining the time series which are the only means we have of assessing whether change has taken place in our environment and, if so, for attributing it to cause – natural or anthropogenic – or for estimating the complex effects of physical and chemical change on the ecosystem. And these of course are very close to the goals that Hjort's ICES had in setting up *Annales Biologiques* in the first place.

We wish them all a long and productive retirement, although, in the case of one or two of them, one is obliged to add the rider "when they decide to take it"!

We have one last question to deal with: that of assessing whether the format of the "Decadal Symposium" can act as an adequate substitute for the aims and intentions of the *Annales*. Here, it is the user who will decide the issue, and we can do no better than close with the sentiments with which the original Editor, Harald Blegvad, introduced the first volume of the *Annales* series (though with "symposium" substituted for "volume"): "This volume is now issued as a *trial one* only – an experiment to solve the problem of form in different ways. In its present shape I trust it may serve as a foundation on which it will be possible to build or – if necessary – to rebuild future issues."

Finally, I have a personal acknowledgment to make regarding the very great part played by my four able Co-conveners – Pentti Mälkki, Günther Radach, Roald Sætre, and Michael Sissenwine – in organizing, exhorting, checking, and reviewing the large number of texts that were submitted and also, in the case of Pentti Mälkki, for arranging such a beautiful and appropriately scaled venue as the Ålands Parlamentshus.

We are accustomed to seeing such acknowledgments in print of course and some cases are doubtless better deserved than others. However, in the case of the present Symposium and Festschrift, the Reader should be in no doubt that my own illness and convalescence in the summer and autumn of 1990 meant that these four individuals bore most of the workload themselves throughout the crucial period in the Symposium's development; the successful outcome was therefore largely the result of their hard work – entirely in keeping with the type of individual effort which has motivated the subject in ICES since its beginnings and which I have tried to underscore and acknowledge throughout these introductory remarks.

R. R. Dickson
Convener

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The following is not, of course, intended as a complete reference list for the five main recipients of this Festschrift, nor is it even intended to cover their most important works. It merely lists those papers which it was necessary to describe in composing and illustrating the Introduction.

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Climate and hydrographic variability in the ICES area during the 1980s

D. J. Ellett and J. Blindheim

Ellett, D. J., and Blindheim, J. 1992. Climate and hydrographic variability in the ICES area during the 1980s. – ICES mar. Sci. Symp., 195: 11–31.

The paper notes some global climate research with relevance to the ICES area and reviews advances during the 1980s in the concepts of the circulation of the North Atlantic and the Nordic seas. Published indications of regional changes during the decade are summarized and work in progress from major projects during the period is noted. On present evidence, the 1980s appear to have been a rather "normal" period, with the eastern and northern parts of the area recovering from the low temperature and salinity levels of the Great Salinity Anomaly in the earlier years, and with a less cold, fresher period in the northwestern Atlantic in the first half of the decade which may be tentatively linked with fresher conditions in the eastern Atlantic during the second half. The Subpolar Gyre remained rather cool by comparison with the years before 1970.

D. J. Ellett: Dunstaffnage Marine Laboratory, PO Box 3, Oban, Argyll PA34 4JR, Scotland. J. Blindheim: Havforskningsinstituttet, PO Box 1870, N–5024 Bergen-Nordnes, Norway.

Introduction

The first section of this symposium volume offers detailed reviews of the 1980s' hydrography for most of the main regions of interest to ICES workers. So what should an introductory paper do? Three possible aims are: to examine the oceanographic implications of recent research upon climate, to review current concepts of the circulation of the North Atlantic and the Nordic Seas, and to note the major programmes of the decade and any published results. Because these topics are interdependent, we shall not make a rigid attempt to keep them separate.

Major temperature changes during the century

The possible consequences of global warming have meant that much recent research on the climate has been devoted to establishing reliable values for the magnitude of surface air temperature changes over the past century. Because of the predominance of the oceans upon the Earth's surface, analysis of sea surface temperature (SST) records has formed an important part of this (e.g. Folland *et al.*, 1984: Jones *et al.*, 1986). Of relevance to the use of SSTs from meteorological databases for oceanographical purposes is the finding that corrections are required as a result of changes in the methods of observation. For instance, Jones *et al.* (1986) suggest that SSTs from the last century were too cool for consistency with air temperature changes from the same sources by 0.08°K, those from 1903–1941 require a correction of +0.49°K and any from 1942–1945 are too warm by 0.10°K. Subsequent SST data need no correction. Thus, although the topic of comparability between meteorological SSTs and oceanographical data was reviewed by ICES during the preparation of the North Sea temperature and salinity atlas (ICES, 1962), it appears that much care is needed in the use of older data. The final series of annual mean surface temperature variation for the Northern Hemisphere from 1861–1984 of Jones *et al.* (1986) is shown in Figure 1 as anomalies from the mean period 1950–1979 and was formed using 50% weightings for land temperature and corrected SSTs. An overall rise of about 0.6°K is apparent. The largest rise occurred between about 1917 and 1938, with a subsequent colder period reaching its minimum during the 1970s; 1981 and 1983 were the warmest

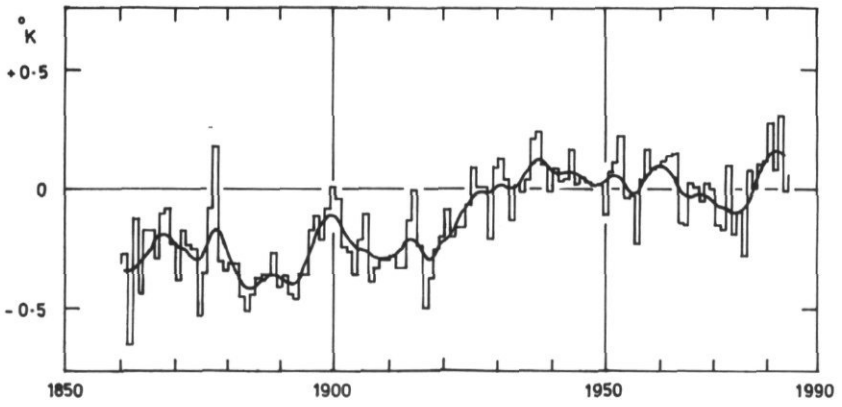

Figure 1. Northern Hemisphere annual mean surface temperature variations (Jones *et al.*, 1986). Smooth curve shows 10-year Gaussian filtered values.

individual years of this series to 1984. (Subsequent years have also been very warm: Globally the six hottest years of the 1859–1990 series have occurred in the last nine years [Ratcliffe, 1991].) This graph and its companions for the Southern Hemisphere and complete Globe (Jones *et al.*, 1986) are closely similar to series compiled by a number of authors and can be regarded as the present best estimates for surface air temperature changes during the past century; further analyses will be discussed later. However, we can expect that ocean surface temperatures will differ regionally, if only because of thermal lag. Figure 2 shows temperature anomalies of 1947–1986 with respect to 1951–1980 means for the area 40°–70°N, 10°–60°W (Jones *et al.*, 1987), using SST and land data weighted by area. It shows a maximum in the early 1950s followed by a decline to the early 1970s. Following a secondary maximum around 1980, temperatures fell until the end of the series to values about 0.4°K below the 1951–1980 mean values. The cool North Atlantic of recent years therefore contrasts with the warming air temperatures of Figure 1. Between 1947 and 1977 Figure 2 can be compared with the Smed series of SST anomalies (Smed *et al.*, 1982) shown in Figure 3, where it can be seen to resemble the curve for Smed's area E, suggesting that the values of Jones *et al.* are weighted towards the south and west of their rectangle where there is a greater volume of shipping traffic. Smed's Northeast Atlantic anomalies showed a temperature maximum towards the end of the 1950s, whilst those around Greenland peaked in 1960–1965.

New views of the hydrography of the ICES Area

The North Atlantic subpolar gyre

The progressive lag, noted above, of the 1950s temperature maximum was one of the factors directing attention during the last decade back from the dense northern outflows which had claimed attention in preceding years towards the horizontal gyres of the northern Atlantic. During the 1980s further interrelationships between parts of the North Atlantic Current system were developed from the concepts which Dietrich *et al.* (1975) based upon the results of the IGY Polar Front Survey. McCartney and Talley showed (1982) that the water circulating in the North Atlantic Subpolar Gyre can be regarded as a mode water, which in its passage around the gyre (McCartney and Talley, 1984) is converted from a warm, saline water mass to a cold, fresher one by its contact with the atmosphere and adjacent waters. This also highlighted the role of the Labrador Sea (Talley and McCartney, 1982) as a sink for a significant

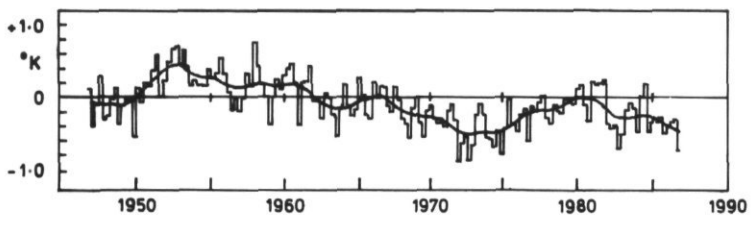

Figure 2. North Atlantic seasonal surface temperature anomalies from 1951–1986 means (Jones *et al.*, 1987) formed from SST's and land data weighted by area. Smooth curves show 5-year Gaussian filtered values.

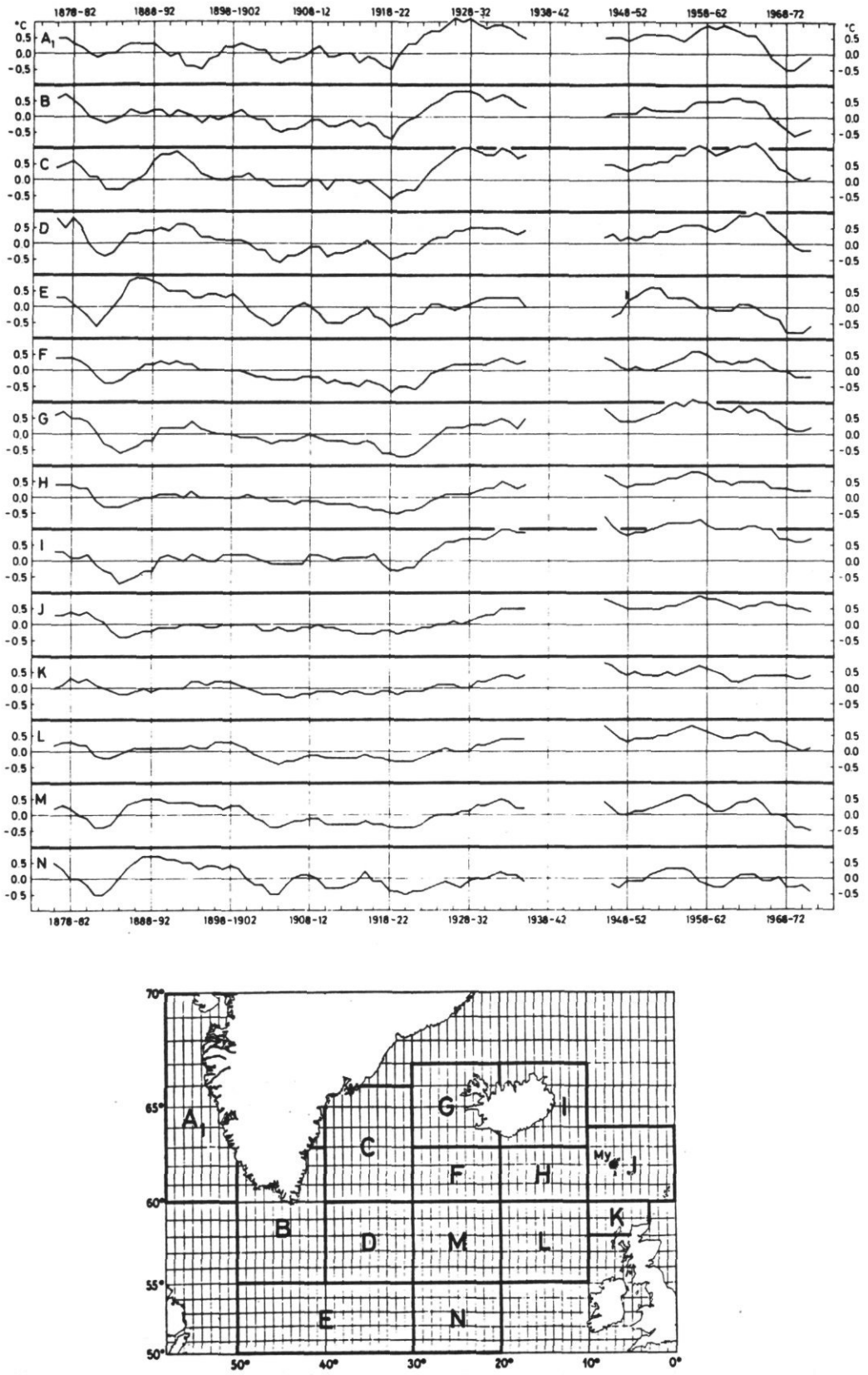

Figure 3. Five-year running means of SST anomalies from 1876–1915 means (Smed *et al.*, 1982), and location of areas.

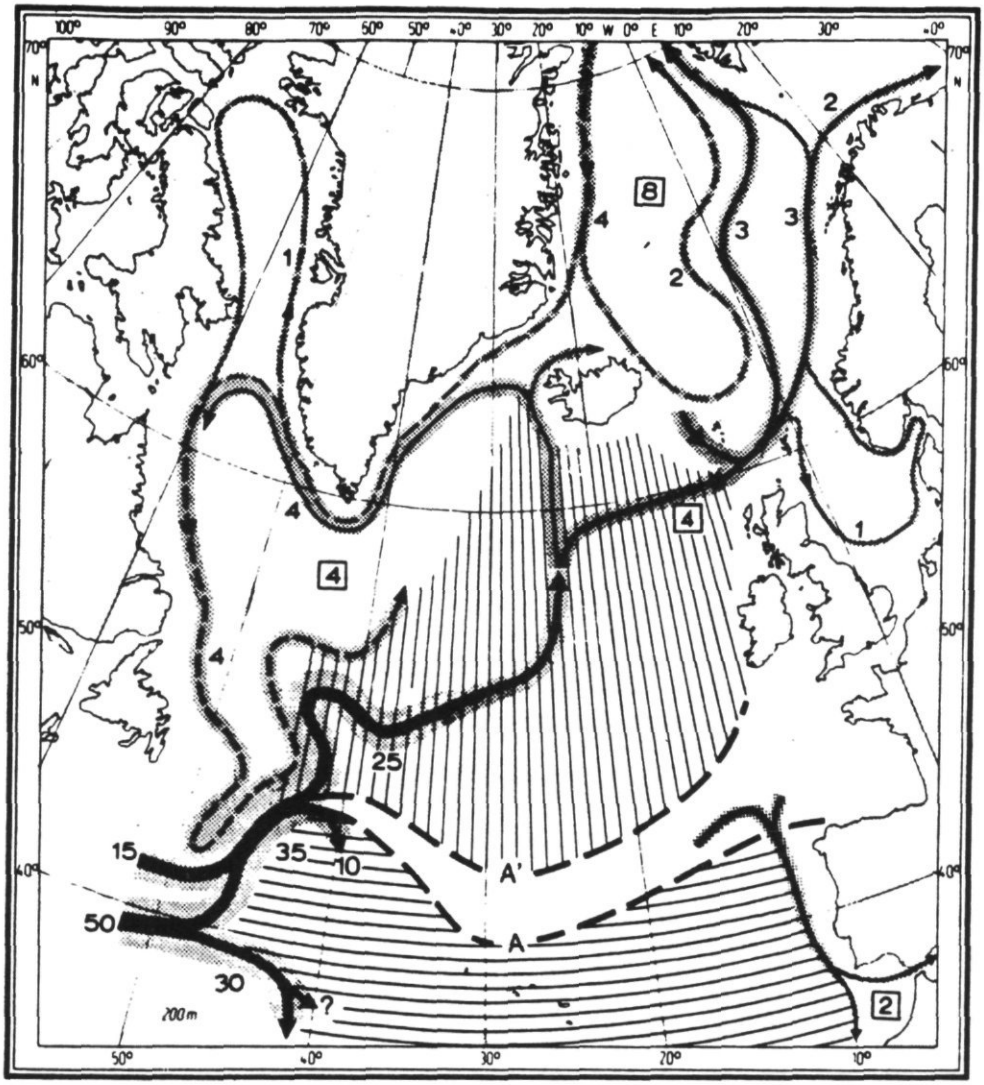

Figure 4. Scheme of North Atlantic circulation as modified by Krauss (1986) from the scheme of Dietrich *et al.* (1975). Numbers represent volume transport in 10^6 m^3 s^{-1}.

part of the North Atlantic Current water returning northwestwards from the Northeast Atlantic, helping to account for disparities between measured inflow to the Norwegian Sea from the Atlantic and the larger flow and heat loss requirements of Worthington's (1970) box model. Nevertheless, during the 1980s the sources of Norwegian Sea inflow have received scrutiny, and in addition to the flow over the deep water between Scotland and Rockall, long thought of as the principal route (Nansen, 1913), input by the slope current west of Britain (Booth and Ellett, 1983) and across the Faroe–Iceland Ridge (Hansen *et al.*, 1986) has been shown to be significant.

Ninety satellite-tracked drogues were used in 1981–1984 by Krauss (1986) to study the branching of the North Atlantic Current between the Newfoundland Grand Banks and the Mid-Atlantic Ridge. The permanent branches of Dietrich *et al.* (1975) were not found, much of the transport of the gyre being in a broad easterly flow in about 50°N, parallel to the Polar Front (Fig. 4). A lesser easterly flow centred on about 40°N formed the northern part of the Subtropical Gyre and carried water to the Portugal Current. The Topogulf experiment of 1983–1984 investigated flow across the Mid-Atlantic Ridge within six boxes between 24° and 53°N; Sy (1988) reports rather similar findings, except that the northern flow occurred in two variable branches, and the southern flow was in 32°–35°N. Topogulf water masses have been analysed by Harvey and Arhan (1988) and Arhan (1990) and clarify the Mid-Atlantic extent of Subarctic Intermediate water, which is subducted at the Polar Front. Pollard and Pu (1985)

have documented the sinking of surface water from the northeastern quadrant of the Subtropical Gyre between the Mid-Atlantic Ridge and Iberia. With the addition of eastern boundary currents (e.g. Huthnance and Gould, 1989) and flow across the Faroe–Iceland Ridge (Hansen *et al.*, 1986), the scheme of Krauss (1986) (Fig. 4) makes a useful working summary of our present knowledge of the Subpolar Gyre.

The Norwegian and Greenland seas

In the Norwegian and Greenland seas also, concepts have become modified during recent decades in the light of new research. As mentioned above, incoming Atlantic water may cross the Scotland–Greenland Ridges by a number of routes, but in the eastern sector it is recognized that recirculation in the Faroe–Shetland Channel (Dooley and Meincke, 1981) brings together the Atlantic waters of the slope current, Rockall Trough, and Faroe–Iceland Ridge into the appearance of a common source from this Channel. Hansen and Jákupsstovu (this volume) discuss patterns of circulation here during the decade.

Major advances in the knowledge of the origin and formation of the deep and intermediate water masses of the northern seas have come from the ICES Deep-Water Project (Clarke, ed., 1990). One of the results of the project (Clarke *et al.*, 1990) is of particular interest in relation to climatic fluctuations. Examination of deep temperature data from the central Greenland Sea over the last three decades demonstrates that Greenland Sea Deep Water (GSDW) properties are variable with time. Previously, Aagaard (1968) had shown evidence of a warming between the beginning of the century and the 1950s. Temperatures below 2000 m compiled by Clarke *et al.* (Fig. 5) are consistent with Aagaard's results during the 1960s, and show a warming trend in the early 1980s.

The Deep-Water Project data set also revealed that the Norwegian Sea Deep Water (NSDW) which fills the Norwegian Sea basin below about 2000 m depth is composed of approximately equal parts of GSDW and Eurasian Basin Deep Water (EBDW) from the Arctic Ocean (Aagaard *et al.*, 1985; Swift and Koltermann, 1988). Previously, salt flux from the overlying Atlantic water in the Norwegian Sea was thought to be responsible for converting GSDW to NSDW, but Blindheim (1990) has shown that a widespread intermediate layer prevents this. The same water mass has recently been identified in the Faroe–Shetland Channel, where Martin (1988) links its salinity with the salinity of Atlantic inflow to the Norwegian Sea about seven years previously.

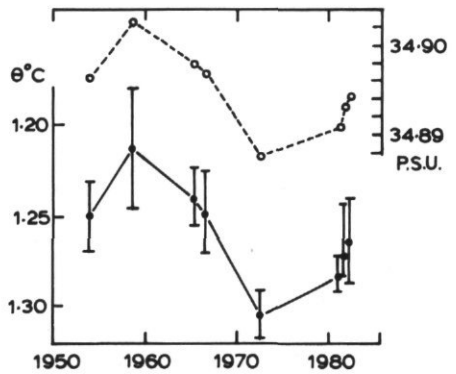

Figure 5. Potential temperature changes below 2000 m in the Greenland Basin and salinity changes deduced assuming constant density (from Clarke *et al.*, 1990).

The Barents Sea

The relevance of the Barents Sea to the climate system of the Arctic Ocean and North Atlantic has become clearer during the 1980s. Aagaard *et al.* (1985) postulated that the source of the high-salinity EBDW of the Arctic Ocean was from the surrounding Arctic shelf, and it has become evident that the northern and north-eastern Barents Sea is an important area in this context. On Novaya Zemlya Bank and Central Bank, Midttun (1985) found dense bottom water with relatively high salinity, resulting from mixing with dense water from ice formation. Dynamic topography indicated that this water was flowing eastwards towards the St Anna Trough, with a density sufficient to allow it to reach great depths in the Arctic Ocean.

In the western Barents Sea, Midttun also observed dense bottom water in the Storfjord area of the Svalbard shelf; this was further studied by Quadfasel *et al.* (1988 and this volume), who were able to trace the flow of this water to more than 2000 m depth in the Fram Strait. The dense plume was followed along the western continental slope of Svalbard from 76° to 79°N. These Arctic inputs are necessarily balanced by part of the surface Atlantic inflow to the Barents Sea.

Changing perceptions of oceanographic variability

The improved accuracy and resolution of oceanographic observations since the later 1950s allowed significant changes in the deep ocean to be noted during the early 1980s as a result of the repetition of sections worked in the IGY surveys of 1957–1959 (Roemmich and Wunsch, 1984) and the detailed sections of the Transient Tracers in the Ocean (TTO) study (Swift, 1984). The recent completion of an atlas of the North Atlantic for the early 1980s (Fukumori *et al.*, 1991) is a preliminary to further analysis of changes in the course of 20 years, and although strictly outwith the remit of this volume, the extensive analysis of the existing database for interpentadal variability between 1955–1959 and 1970–1974 by Levitus (1989a, b and c) should be noted here.

With increasing emphasis on process studies in the ocean, ideas about variability have become further modified, and new parameters of change have come into use. Long-term current meter records have shown the deep-sea to be a more irregular environment than previously supposed (e.g. Gould and Cutler, 1980) and statistics of eddy kinetic energy have shown both seasonal cycles and interannual differences (Dickson *et al.*, 1986). A compilation of available records (Dickson, 1990) usefully picks out in summary the high- and low-energy areas of the Atlantic. In other studies benthic "storms" with short period velocities of up to 27 cm s^{-1} were observed at depths exceeding 4000 m (Klein, 1987). Warm-core (Le Groupe Tourbillion, 1983) and cold-core (Mittelstaedt, 1987) eddies have been studied in the northeastern Atlantic during the 1980s, and in these cases the former displayed connections with Mediterranean outflow, and the latter with the Polar Front. Such sources of variation show the need for caution in making comparisons at oceanic sites where sampling is infrequent.

Large-scale connections and feedback across the ICES Area

The "Great Salinity Anomaly" and its aftermath

No single event demonstrated the reality of the Subpolar Gyres of the Atlantic and Nordic Seas more clearly than the "Great Salinity Anomaly", documented by Dickson *et al.* (1988a) in its transit from East Greenland in the late 1960s, via the Labrador Sea, Northeast Atlantic and Norwegian Sea back to East Greenland in the early 1980s (Fig. 6). Apart from the reduction in surface salinity, other processes were affected; north of Iceland, salinity fell below the critical value for ice formation (Malmberg, 1969); in the Labrador Sea, increased stability of the surface layer severely reduced winter convection (Blindheim, 1967, 1974; Lazier, 1980) and by cutting off annual renewal of the deep layers allowed deep salinity values to increase; the fresher water available to the northwest of the oceanic polar front allowed notably colder and fresher Subarctic Intermediate water to reach Ocean Weather Station Juliett (Ellett, 1982) and the Arctic Intermediate water of the Faroe–Shetland Channel had lower salinity as a result of fresher water crossing the Faroe–Iceland Ridge (Martin, 1981). These are relevant points, as in some areas effects have been observed to persist well into the early 1980s.

The "Great Salinity Anomaly" and its associated effects have also stimulated the examination of the role of salinity in past and future climates. Recent work reviewed by Street-Perrott and Perrott (1990) has recognized that past Atlantic Ocean surface stability variations may have given rise to sudden changes resulting in such diverse phenomena as periods of tropical drought and discontinuities in deglaciation.

Ice and large-scale hydrography

The interrelationships between climate, ice, and the oceanography of the polar seas are too obvious to need stating. However, as wider links and feedback mechanisms in the North Atlantic gyres are established, it becomes necessary to be aware of variations in the remotest parts of the system. This was apparent as the case study of the 1970s salinity anomaly was carried out (Dickson *et al.*, 1988a), and implications continue to emerge. Mysak *et al.* (1990) have investigated sea-ice anomalies in five regions between the Fram Strait and the Labrador Sea and their connection with Atlantic salinity anomalies. They find that positive sea-ice anomalies propagated southward during 1968–1973, contemporaneously with the earliest observations of the Great Salinity Anomaly, and were apparently influenced by extreme runoff into the western Arctic basin from northern Canada and Alaska during 1964–1966. The larger extent of the sea-ice is believed to result from the suppression of convection by the low salinity of the surface layer, rather than the low salinity resulting from the ice anomaly. Mysak *et al.* (1990) propose a 10-component negative feedback loop connecting Arctic cyclogenesis, surface salinity, sea ice, and Atlantic/Arctic advection (Fig. 7). This might reverse upon an interdecadal time scale of roughly 20–30 years, and they quote some preliminary data for positive ice anomalies in the Greenland Sea during the winters of 1987–1989, which with the evidence of earlier anomalies helps to support the proposed time scale of events. They also note that the cycle would be non-linear, with rapid effects from runoff and sea-ice extent, but with slow return to initial conditions, as observed following the "Great Salinity Anomaly".

Ådlandsvik and Loeng (in press) propose a similar, but abbreviated, feedback cycle for the Barents Sea climatic system. Variations in the Atlantic inflow determine the position of the Barents Sea polar front, which in turn influences the winter ice coverage, and hence the heat flux to the atmosphere. Ådlandsvik's (1989) model shows the importance of regional atmospheric forcing upon flow variations. The period 1970–1986 has been modelled, and although a standard external inflow is assumed, internal fluxes resulting from local wind forcing produce good agreement in temperature, ice and air pressure fields. Relatively low pressure over the Barents Sea during the first half of the 1970s was followed by a high pressure period during 1976–1980, low pressure during 1982–1983, and a return to high pressure during 1984–1985. Computed flux through the Fugløya–Bjornøya section follows this general pattern (Fig. 7, Loeng *et al.*, this volume), with low pressure increasing the flow. It also correlates very closely with observed

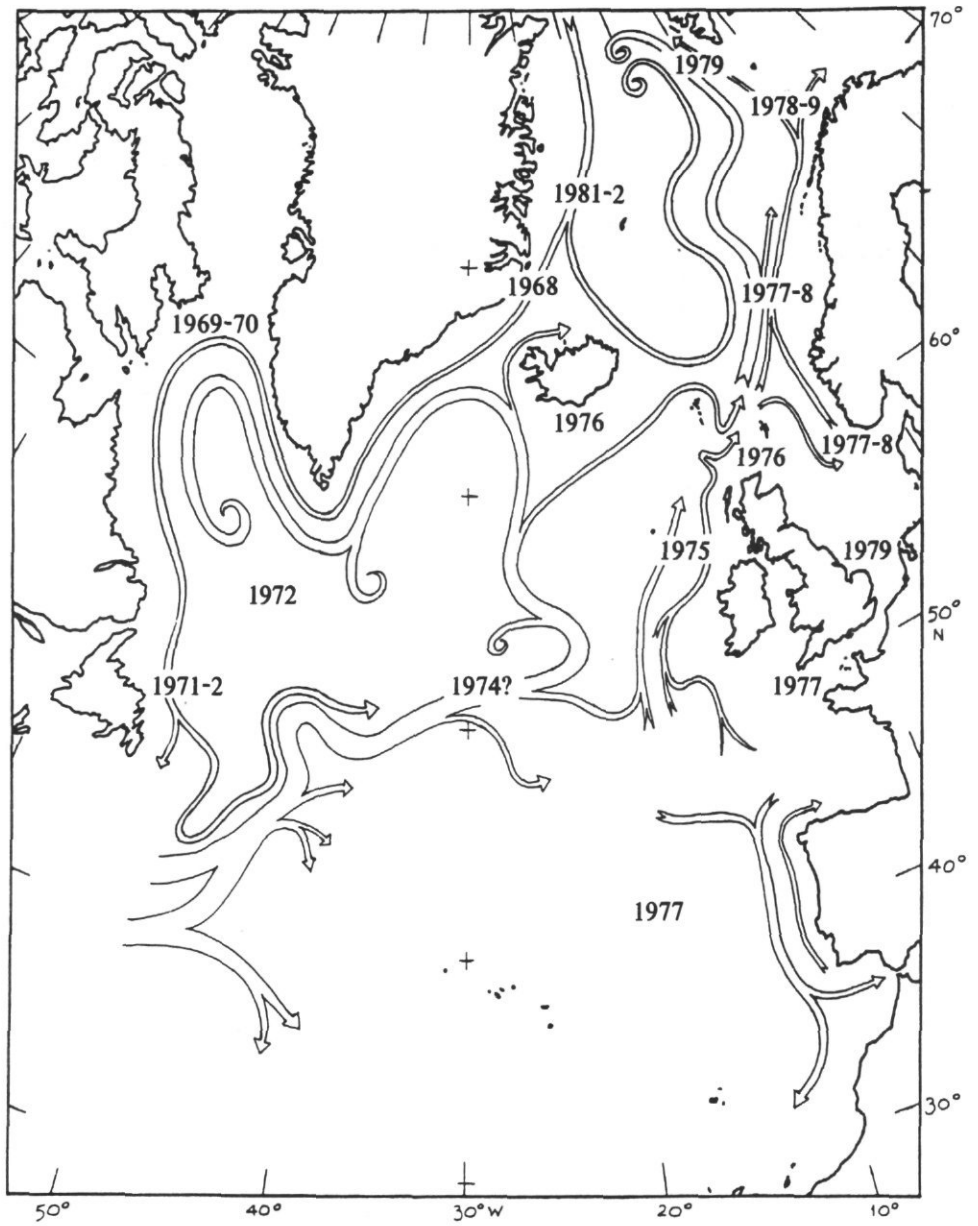

Figure 6. Transit dates for the minimal values of the "Great Salinity Anomaly" of Dickson *et al.* (1988). The Krauss scheme of circulation has been modified in the NE Atlantic after Meincke (1986) and others.

temperature anomalies on the Kola meridian (Bochkov, 1982), which against the 60-year mean show an alternation of warm periods in 1972–1976 and 1982–1984 with cold periods in 1971, 1977–1982, and 1985–1986. The ice index for the Barents Sea (Loeng, 1979), which shows increased ice cover during 1978–1982, also corresponds in detail with the modelled fluxes. The study leads Ådlandsvik and Loeng to suggest that the Barents Sea climate oscillates between a warm and a cold state through the proposed feedback mechanism, and they suggest that the formation of bottom water during the winter is important in determining the state. If dense bottom water leaves the northeastern Barents Sea to enter the interior of the Eurasian Basin of the Arctic Ocean, as occurred in 1982–1983 (Midttun and Loeng, 1987), a compensatory inflow of Atlantic water may occur. Alternatively, increased Atlantic inflow may displace bottom water from the Sea. Ådlandsvik and Loeng prefer the second hypothesis.

In some regions large changes at the end of the ice season have been noted. Vinje and Kvambekk (in press) have digitized weekly ice distributions for 1966–1988 for

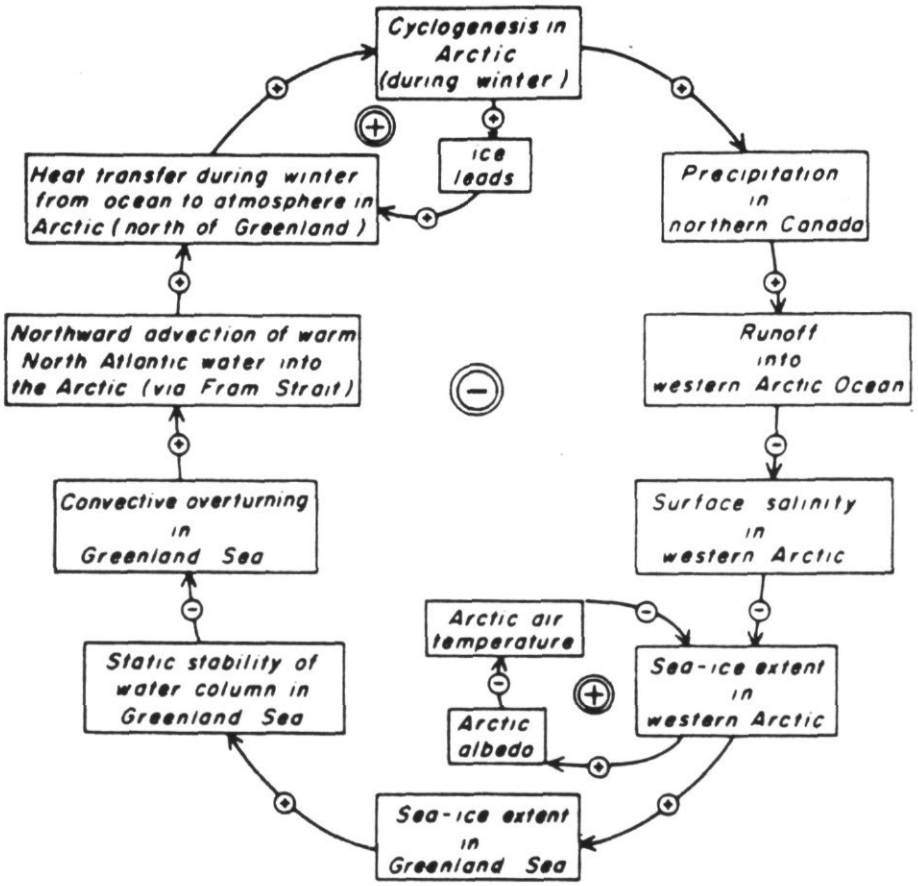

Figure 7. A negative feedback loop for an interdecadal Arctic climate cycle as suggested by Mysak *et al.* (1990).

the Barents Sea, and find that over this period in the area between 68°–83°N and 10°–55°E ice cover at the end of the melting season was reduced by 40%, although the annual maximum extent was not reduced. They connect this with reduced ice thickness or more rapid melting processes over the 23-year period. Vinje (1990) has reported that in the Iceland Sea both the average minimum and maximum extensions showed a decrease over the same period since the severe Icelandic ice years of the late 1960s by 40 and 33% respectively, but he found no similar reduction in the Greenland Sea. Gloersen and Campbell (1991) use two-day satellite cover of the poles throughout 1978–1987 to show a 2% reduction in Arctic ice cover during this period. Contemporaneous Antarctic ice cover changes were not significant.

Cycles and global warming?

Practical oceanographers, experiencing the unfolding of ever more detailed feedback mechanisms, have distrusted most theories of long-term cycles. However, the hypotheses above show that lengthening records are helping to produce realistic models for testing. In the Southern Hemisphere the El Niño–Southern Oscillation (ENSO) clearly demonstrates major cyclic ocean-atmosphere coupling, although of varying frequency and intensity. The North Atlantic Oscillation (NAO) proposed with the Southern Oscillation by Sir Gilbert Walker in the 1920s is less well documented, except in the inverse associations between mean winter temperatures in Greenland and Northern Europe and between SST east of Newfoundland and pressure over the north-eastern Atlantic (e.g. Lamb and Ratcliffe, 1972). Rodionov and Krovnin (this volume) show that effects of the NAO were apparent during the 1980s. It is perhaps appropriate here to note an interesting analysis of the Jones *et al.* (1986) global temperature variation curves, of which we have shown the Northern Hemisphere graph in Figure 1. Recently, Ghil and Vautard (1991) have further analysed these data series with a view to separating the warming trend from natural variability. Using singular spectrum analysis, they find that the global trend increased by 0.4°K from 1910 to 1988, with interdecadal oscillations of 21 and 16 years and interannual oscillations of 6 and 5 years. They suggest that the shorter periodicities reflect global effects of ENSO phenomena, and that the longer interdecadal period is connected with extra-tropical watermass variations be-

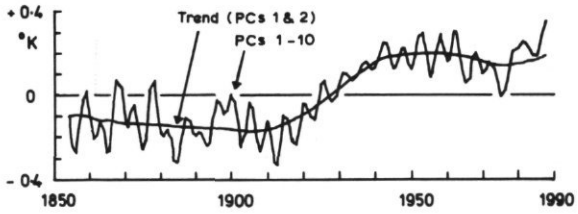

Figure 8. Reconstruction of the temperature curve of Figure 1 by Ghil and Vautard (1991) from an underlying trend and interdecadal oscillations obtained using singular spectrum analysis.

low the seasonal thermocline. Reconstruction of the temperature series from the principal components (Fig. 8, for the Northern Hemisphere) suggests that the bidecadal oscillation was near to maximum in 1988 as was the larger ENSO-related component. Ghil and Vautard therefore suggest that a global temperature decrease is probable over the next decade, despite the increase in the underlying trend.

Climate during the 1980s

Mean pressure patterns over the Atlantic and Northern seas

Figures 9a–e show the differences of pressure between means for 1980–1988 and 1951–1980, drawn from values at 5° latitude ×10° longitude intersections computed by the Climatic Research Unit, University of East Anglia (P. M. Kelly, pers. comm.). The most notable feature of all five diagrams is the lower 1980s' pressure over Greenland. This difference, which exceeds 6 mb in the winter and spring distributions, shows a return to near-normal conditions from the persistent Greenland high pressure anomaly of the 1960s (Dickson *et al.*, 1988a) which strengthened northerly winds over the East Greenland Current. Elsewhere, annual differences in comparison with the previous 20 years are unremarkable, with a weak west-to-east anomaly wind over the Atlantic Current and a weak south-to-north anomaly wind over Scandinavia. A 3 mb anomaly high in spring in 50°N, 40°W is counterbalanced in autumn by a 2 mb low in the same latitude. Only over the Labrador and Irminger Seas in winter is a stronger anomaly gradient present, and this is likely to represent a decrease of northerly winds with the decline of the Greenland high of the earlier period, rather than an increase of southerlies. Overall, these diagrams suggest that the 1980s was a rather "normal" period, though a discussion of regional variations will show some exceptions. Rodionov and Krovnin (this volume) find evidence of increased interannual variability during the decade, and Dickson, Colebrook and Svendsen (this volume) show changes in summer pressure patterns between the early and later parts of the decade.

Some regional climate changes

Bakun (1990) has investigated long-term trends in wind stress and finds evidence that alongshore stresses have intensified in major upwelling regions over the past four decades. He suggests this arises as a consequence of pressure changes due to contrasts between warming continents and still cool oceans (cf. Figs. 1 and 2). In later work he examines this effect during April to September at a number of locations around the North Atlantic, and finds significant alongshore wind stress increases at Cape Hatteras, off the Iberian Peninsula, and off Northwest Africa (Bakun, this volume, Fig. 4). Dickson *et al.* (1988b) also suggest an intensification of upwelling off Iberia and give evidence of enhanced northerly winds over western Europe between 1959 and 1980, with which they associate a decline in plankton biomass.

On the coast itself, climate may reflect oceanic properties more than continental ones. An example is the annual mean air temperature series for NW Scotland shown by Parker and Folland (1988). A regression line fitted to the 1951–1987 temperature anomalies gives a fall of about 0.5°C over this period, which they ascribe to the decline in mid-latitude North Atlantic SST. Parker and Folland show the global distribution of SST changes for 1971–1987 relative to 1951–1960, which demonstrates that the greatest cooling was centred upon the northwest sector of the Atlantic Subpolar Gyre.

Air temperatures along the Norwegian coast show rather similar variations. Annual temperatures reached a maximum in the late 1940s and were still declining from this peak during the 1980s (Aune, 1989). However, as Figure 10 shows, in the south, from Utsira and Bergen to Trondheim, autumn temperatures have remained above the long-term averages, and in the north spring temperatures have either remained close to the 1940s value (Trondheim) or have risen steadily since the 1880s (Tromsø).

An unusual aspect of climatic change in recent decades has been a tendency towards increased wave height in the North Atlantic. Both visual estimations (Neu, 1984) and instrumental observations (Carter and Draper, 1988) show this feature, which in the latter case amounted to 0.034 m/yr between 1962 and 1985 at the western entrance to the English Channel. Neu finds 3–7-year fluctuations in addition to a continuous increase in sea state between 1970 and 1982. The former are ascribed to variations in cyclonic activity over the western Atlantic and the latter to a steady increase in the strength of westerlies. Certainly Rodionov and Krovnin (this volume) find that Atlantic westerlies have strengthened during the past two decades, but discussion which follows indicates declining westerlies over southern Britain during the same period. The site of Carter and Draper's observations, the Seven Stones Lightvessel, is however very exposed to Atlantic swells and is likely to have felt the effects of mid-Atlantic weather.

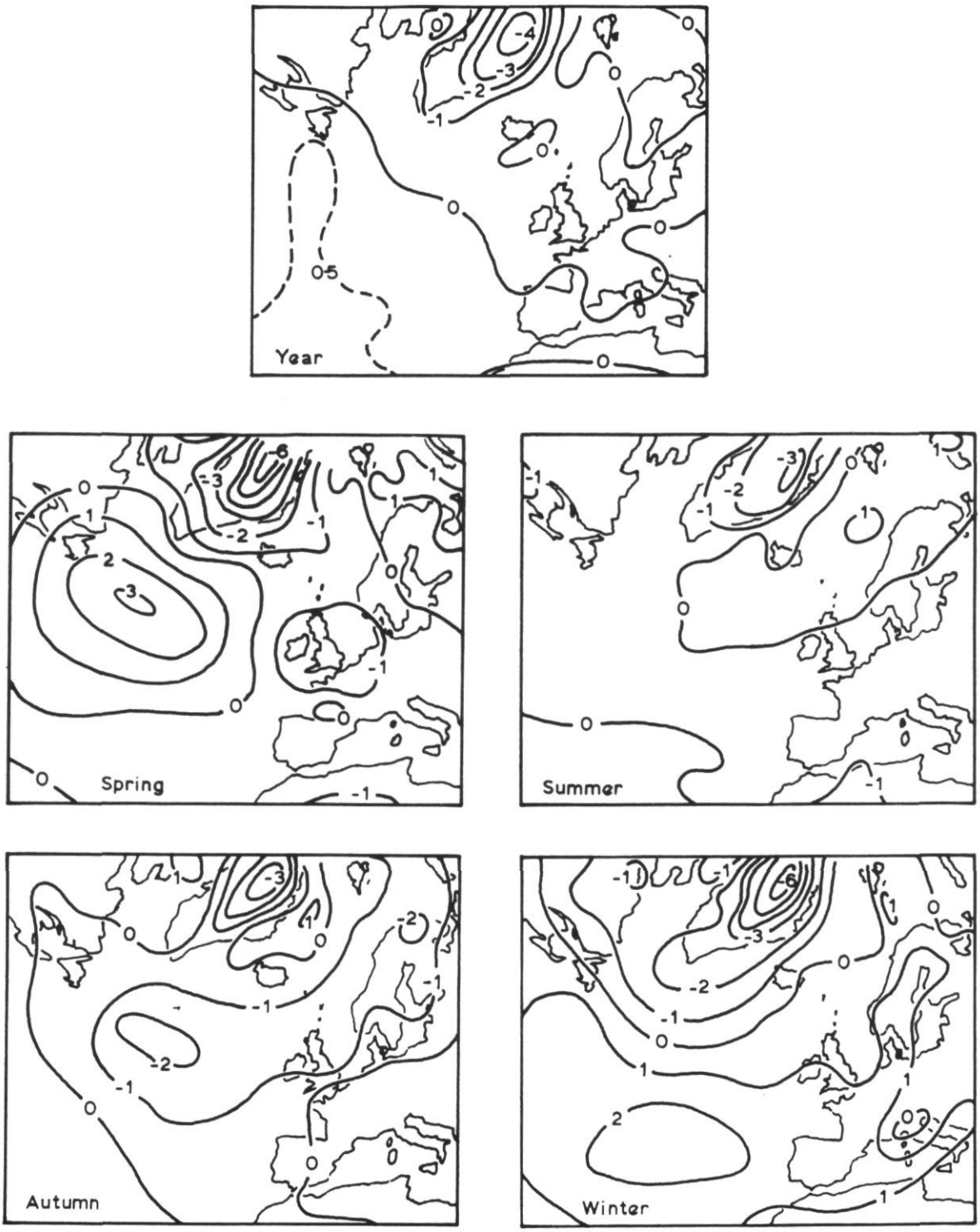

Figure 9. Annual and seasonal pressure anomalies (mb) for 1980–1988 minus 1951–1980 (P. M. Kelly, pers. comm.).

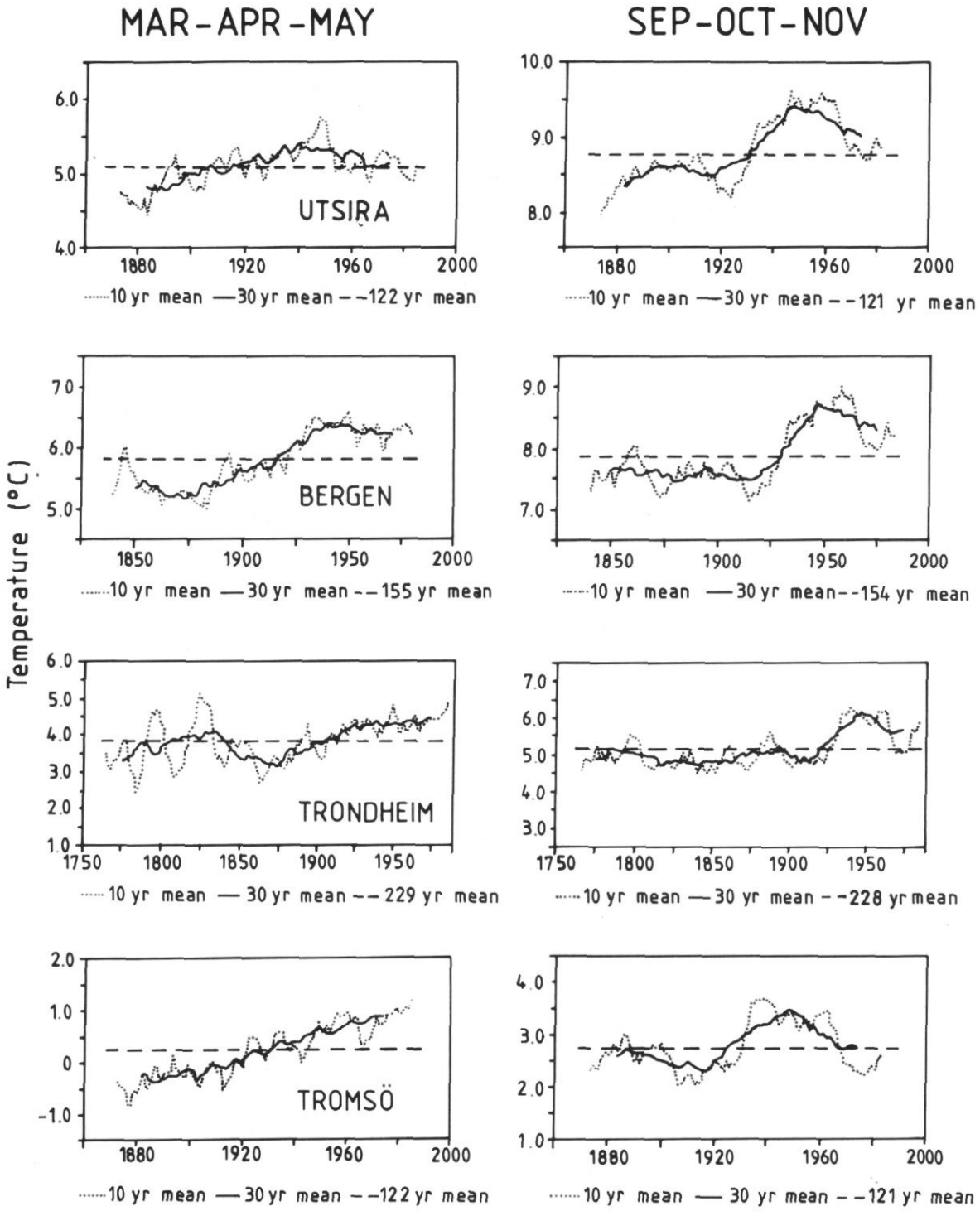

Figure 10. Annual mean air temperatures for spring (Mar–Apr–May) and Autumn (Sep–Oct–Nov) at four stations along the Norwegian coast (Aune, 1989).

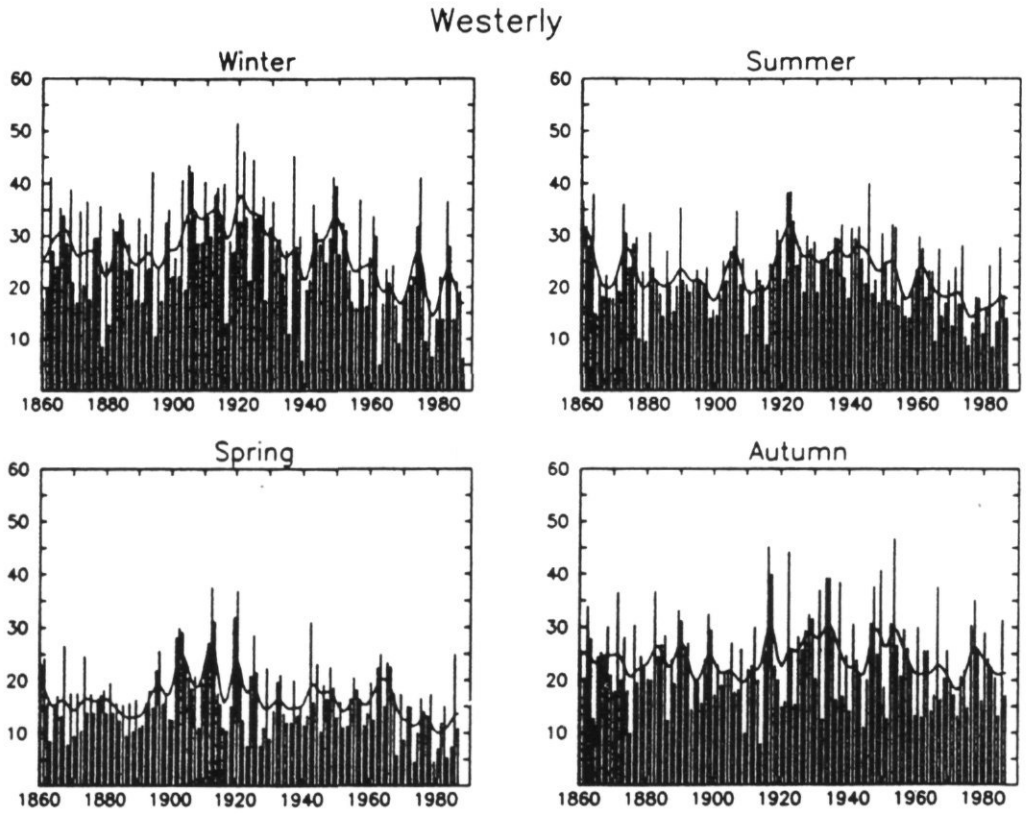

Figure 11. Index (days) of westerly weather over the British Isles, 1860–1987 (Briffa *et al.*, 1990).

A valuable indication of climatic conditions in the vicinity of the British Isles is the Catalogue of Daily Weather Types initiated by Lamb (1972). The series is updated by the Climatic Research Unit of the University of East Anglia and has provided the basis for principal component analyses by Jones and Kelly (1982) and Briffa *et al.* (1990). The most marked feature of the data is the decline in the number of days of westerly weather over the UK since the 1920s, which seems to have continued into at least the early 1980s (Fig. 11). Briffa *et al.* (1990) show that the main compensation for this has been an increase of cyclonicity and anticyclonicity, which was also at the expense of northerly and northwesterly types of weather during the 1980s. In terms of seasons, the westerly decline is greatest in winter and summer, being recently 15 days per winter in comparison with a long-term mean of 25 days. Since 1960, winter easterlies have risen from the long-term mean of 5 days to about 10 days. Autumn has varied least of the seasons. By comparison with earlier decades, in addition to the continuing decrease of westerlies, Briffa *et al.* (1990) show that the 1980–1987 period had annually less frequent northerlies and fewer spring easterlies. Mayes (1991) has noted that regional differences in airflow types can sometimes be important between the southeast and northwest of the UK. From early 1988 to spring 1990 westerliness was greatly increased over Scotland in contrast to increased anticyclonicity in southeast England at this time. As noted earlier, Rodionov and Krovnin (this volume) have found that winter westerlies over the North Atlantic were increasing during the 1970s and 1980s, but their charts of sea level pressure anomaly (their fig. 3) are not inconsistent with fewer westerlies over central England, as indicated by Briffa *et al.*

An investigation into variations of windiness over the UK, irrespective of direction, has been made by Hammond (1990), continuing an index based on anemograph and pressure gradient records begun by Smith (1982). Hammond shows that annual windiness decreased sharply during the early 1980s, giving the lowest values of the index from 1881 (Fig. 12). The number of gales per winter over the UK has been examined by Hulme and Jones (1991). December to February gales were more frequent in the late 1950s to early 1960s than during 1968–1987, though individual years exceeded the mean of 29 days. The winters of 1988–1989 and 1989–1990 showed a return to notably windy conditions.

Climatic variability in the central and northern North Sea is discussed by Svendsen and Magnusson (this volume). Off southwest Norway wind speeds were found to have increased since the mid-1970s, with especially strong spring northerly winds in 1981, 1982, and 1988.

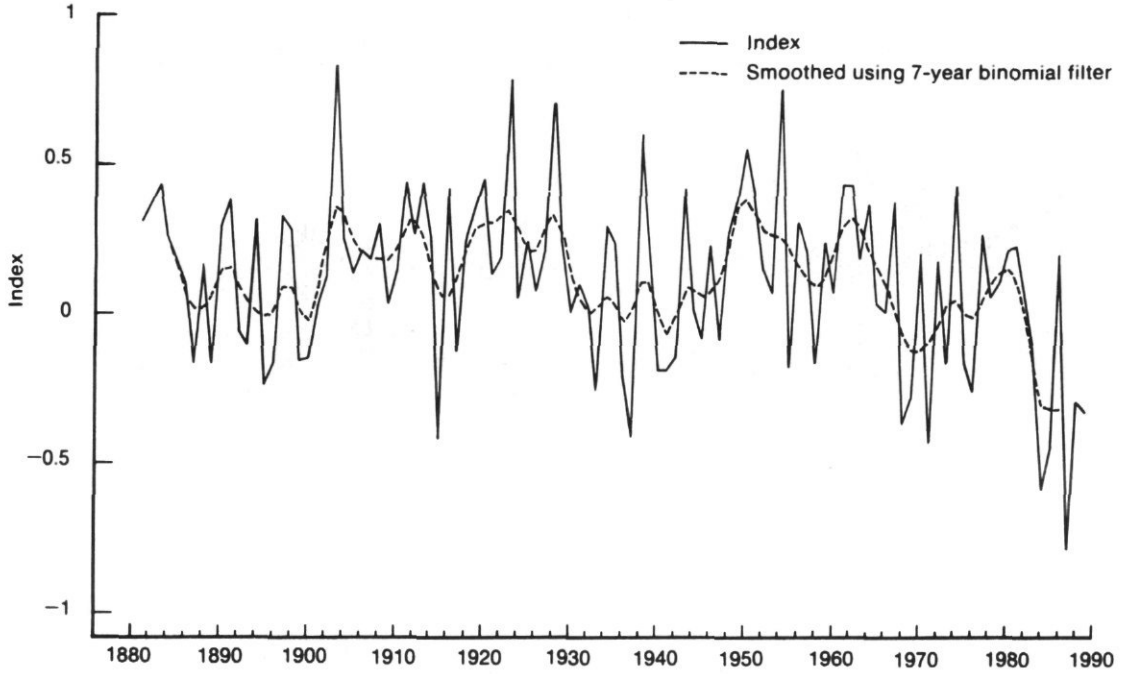

Figure 12. Annual windiness index, 1881–1989 (Hammond, 1990). The dotted line is a 7-year binomial filter.

Regional hydrographic changes of the 1980s

The Northwest and Western Atlantic

Buch and Stein (1989) review West Greenland conditions for 1980–1985. They found cold sea temperatures during 1981–1985, resulting from cold air anomalies and, in 1982, greater than normal inflow from the East Greenland Current. In July 1984 temperature and salinity were both lower than in any July since 1970, and Buch and Stein suggest that the normal inflow from the Irminger Sea was blocked at this time. Stein and Wegner (1990) found a warming of the near-surface layers upon the West Greenland shelf from November 1984 to 1988.

To the south, in the Labrador Current, Lazier (1988a) found cold conditions in the intermediate water at 200 m from the end of 1982 until late 1985. Amounts of ice were large during the winters of 1983–1984 and 1984–1985, and, like Buch and Stein, Lazier ascribes the cold period to cold winters in the eastern Canadian Arctic, suggesting increased ocean heat loss due to mean winds from the continent.

Comprehensive annual reviews of environmental conditions between Cape Farewell and Cape Hatteras have been given for 1982 to 1988 by Trites and Drinkwater (see Trites and Drinkwater, 1990, for all references). The data of greatest interest to ICES workers are those for the Labrador Sea, Flemish Cap, and Gulf Stream, and for these, annual SST anomalies from 1972–1980 means are given in Figure 13. As at West Greenland, 1982–1984 were cold years in the Labrador Sea, but at Flemish Cap 1979–1983 was on the whole a warm period, which was succeeded by cold years in 1984–1986. Over the seventeen years, the Gulf Stream SST varied

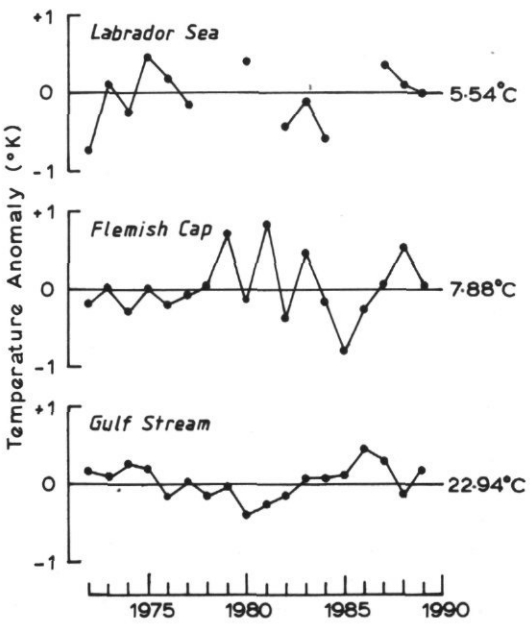

Figure 13. Western Atlantic temperature variations (1972–1989). From data of Trites and Drinkwater (1990).

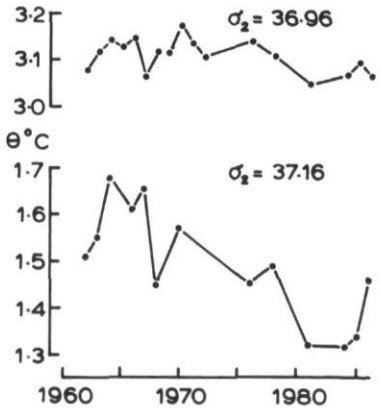

Figure 14. Potential temperatures on two σ_2 surfaces in the western Labrador Sea, 1962–1986 (from Lazier, 1988b).

by only 1°K, below-mean temperatures lasting from 1976–1982, followed by a warm period from 1982–1987.

For the deeper Northwest Atlantic water masses, Wallace and Lazier (1988) encountered the most extreme characteristics observed to date in Labrador Sea Water (Θ = 2.80°C, S = 34.81 ppt) in July 1986, probably formed in the previous winter. Chlorofluoromethane data suggested lack of renewal of Labrador Sea Water (LSW) for some years before 1986. At deeper levels, Lazier (1988b) has observed interannual changes in the characteristics of Northeast Atlantic Deep Water (NEADW) and Denmark Strait Overflow Water (DSOW). Figure 14 shows the 1962–1986 temperature variations on the two potential density anomaly surfaces (referenced to 2000 m, i.e. σ_2) representing these water masses. Changes in the NEADW are small in this region far from its sources in the Northeast Atlantic overflows and dominated, Lazier suggests, by the DSOW variations beneath it. Temperature decreases in 1968 and 1981 may represent the effect of the Great Salinity Anomaly on the overflow water; 1988 temperatures in the DSOW were close to the 1986 value, but in 1990 they had fallen again to about 1.3°C (Lazier, pers. comm.).

The Denmark Strait overflow has itself been a subject of current meter investigations by Dickson *et al.* (1990) between June 1987 and July 1989. They find a flow which fluctuates over short time scales, but which has little seasonal or interannual variation. The overflow across the Denmark Strait sill is estimated at 2.7 Sv, but south of Ångmagssalik this had increased by entrainment and the addition of Northeast Atlantic overflow to 10.7 Sv. The apparent constancy of this quantity and its significance in the deep global ocean will make this a site of continuing interest.

The Mid-Atlantic Ridge and Northeast Atlantic

Although no time-series data for mid-North Atlantic are currently available, we have earlier noted that major German and French work took place in the vicinity of the ridge during the 1980s, much of it under the Topogulf surveys, and these will help to establish values representative of the decade for future comparisons. Data will be available from Ocean Weather Stations Charlie and Lima and from XBT sampling, such as that reported by Fahrbach and Wegner (1987) which was able to show the geographical variability of Subpolar Gyre features. Leach (1990) repeated a section in the upper layers from the Azores to station Charlie in the summers of 1983 and 1986 and noted lower temperature and salinity between 45° and 52°N in 1986 and lesser thermal stratification towards the Polar Front.

Between the Mid-Atlantic Ridge and Iberia, Pollard and Pu (1985) investigated existing data, noting the low salinity of the mid-1970s and have acquired further data during cruises since 1983. Between Lisbon and Madeira, Dias, Amorim and Vacas (this volume) have studied the 1981–1987 surface temperature changes in detail, and Lavín and García (this volume) have analysed the 1980–1989 mean sea level variations along the Spanish Atlantic coast.

Recent results from the western Mediterranean are of interest in connection with exchange at the Straits of Gibraltar. Garrett *et al.* (1990) show that the sea-level fall between the Atlantic and Mediterranean coasts of Spain undergoes rather abrupt changes of 0.1 m about every six months. This may relate to models which suggest that there should be two exchange states; a fast, shallow inflow and a deep, slower inflow. The "flips" could be related to the formation of deep water in the western Mediterranean in winter, which would raise the interface between inflow and outflow. The models imply that maximal exchange occurs with the first state, with the interface shallower than about 90 m. Garrett *et al.* suggest that flips within a month, giving short-term variability, could arise from local meteorological fluctuations.

In the eastern Atlantic, the dynamics of the Mediterranean outflow have remained a topic of investigation during the 1980s. Arhan (1987) uses data from Topogulf to suggest that buoyancy fluxes from salt-fingering beneath the saline plume provide thermohaline forcing to initiate the northward spread of this water in the upper intermediate layers. Zenk and Armi (1990), from 1989 cruises, find an upper core which follows the Portuguese continental slope at about 750 m depth and the main core at about 1250 m. A small portion of this also follows the continental slope, but the major part spreads west and northwest in meanders and eddies, such as that observed by the Tourbillion survey of 1979 (Le Groupe Tourbillion, 1989).

Surface data are available from the Rockall Trough (Fig. 15) from the time series of weather ship and research vessel sampling described for earlier years by Ellett (1982). Because of the deep winter mixing of the area, winter temperature and salinity anomalies carry

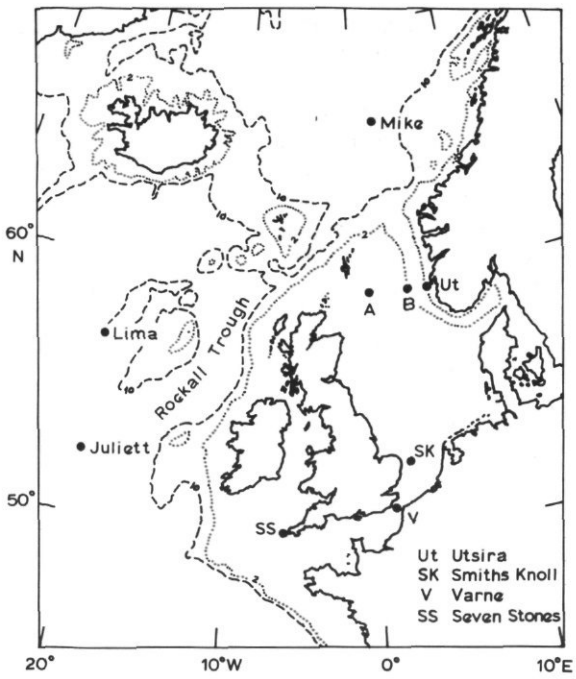

Figure 15. Locations of some eastern Atlantic temperature and salinity series.

most weight, but smoothed monthly salinity anomalies (Fig. 16, including some interpolation) give useful evidence of such events as the Great Salinity Anomaly. This passed through the Rockall Trough in late 1975 to early 1976, but recovery of salinity to near former levels took until 1980. Figure 16 shows a less extreme, but still marked, minimum in summer 1981. It is tempting to suggest that this could be a return via the Irminger Sea of the earlier anomaly, but supporting evidence is at present lacking. Apart from this feature, the early 1980s had relatively high salinity, but were followed in 1985–1989 by a period with values below the 1961–1970 mean. The brief rise in late 1989 to levels close to those of the 1960s has been followed by a return to lower values.

During 1990 a fall in salinity of 0.04 ppt was observed in the lower water column of the central Rockall Trough at depths occupied by Labrador Sea Water. During the 1980s most variations in this water were within 0.02 ppt, and the 1990 fall may be connected with the very low values observed in the area of LSW formation in 1986 by Wallace and Lazier (1988).

Nine-month current meter records of the overflow of Norwegian Sea Deep Water (NSDW) at the Wyville–Thomson Ridge were obtained in 1987–1988. Although a rather steady westward flow was found close to the bottom, in mid-water an eastward current in autumn and winter had periods of strong reversal during spring. This appeared to be a result of the relaxation of the wind field over the ridge crest. In March to June 1988 the mean inflow across the ridge was estimated at 0.3 Sv, containing 25% pure NSDW (Ellett, 1991).

A remarkably steady outflow from the Faroe Bank Channel has also been found by Saunders (1990), who has combined current meter records, CTD observations and Acoustic Doppler current profiles to obtain an estimate of 1.9 Sv for water colder than 3°C entering the Northeast Atlantic by this route. The overall steadiness of these current meter records and those of Dickson *et al.* (1990) from the Denmark Strait noted above is seen as evidence that these outflows are controlled hydraulically by the very large reservoir of the Norwegian and Greenland Seas rather than by local meteorology or short-term variations in inflow to the northern seas.

The English Channel and North Sea

In European coastal waters the pattern of data collection has changed appreciably during the past decade. The automation of light vessels, changing patterns of trade and the move to larger, faster ships has meant the demise of many surface temperature and salinity series of long standing. On the other hand, more intensive research work such as the North Sea Project and associated programmes has led to a much wider coverage of shelf sea processes, and the increasing use of satellite imagery is allowing truly synoptic information to be obtained (Holligan *et al.*, 1989).

Upon the evidence of some of the long-term series which survived into the latter half of the 1980s, conditions in the English Channel and Southern Bight were rather similar to those of the Northeast Atlantic; a little cooler and fresher than in the 1950s and 1960s. At the Seven Stones Lightvessel (LV), between the Scilly Isles and Land's End (Fig. 15), where observations ceased in September 1987, monthly mean temperatures and salinities from January 1980 were as a whole 0.3°K and 0.03 psu below the 1905–1954 means of the ICES (1962) atlas. In the eastern channel, the Varne LV collected data until November 1985, and mean anomalies from monthly means for 1952–1966 (Ellett, 1967) were −0.11°K and −0.09 psu. At the Smiths Knoll LV (52°43′N, 2°18′E), monthly means from January 1980 to October 1988 were 0.29°K and 0.10 psu below the 1952–1966 means. At the Varne and Smiths Knoll LVs, 1982 was the warmest of the observed years, and at the latter station and the Seven Stones, 1986 was the coldest, chiefly due to a cold winter and spring. Pawson (this volume) has English Channel coastal temperature data for the complete decade which confirm 1982 and 1983 (with 1989) as the warmer years, but which show 1980 to have been colder than a group of cold years centred upon 1986. Figure 17 summarizes the mean and extreme values for the Smiths Knoll LV. Data for these positions have kindly been supplied by S. R. Jones, MAFF, Lowestoft (pers. comm.).

The 1980s events in the central and northern North

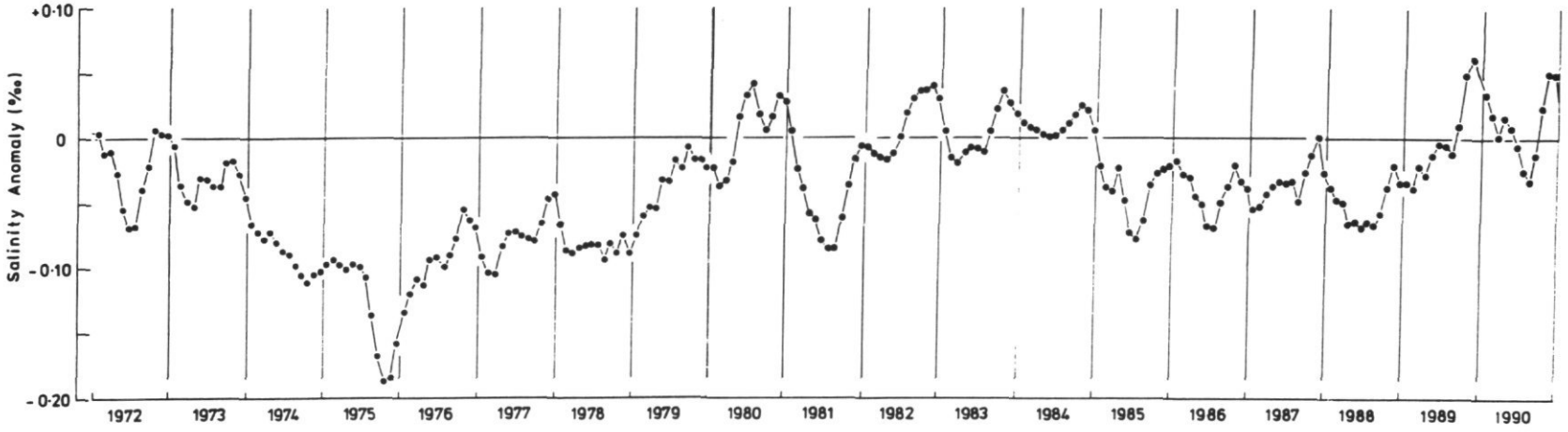

Figure 16. Running three-monthly means of monthly salinity anomalies from 1961–1970 means for the Rockall Trough (1972–1990).

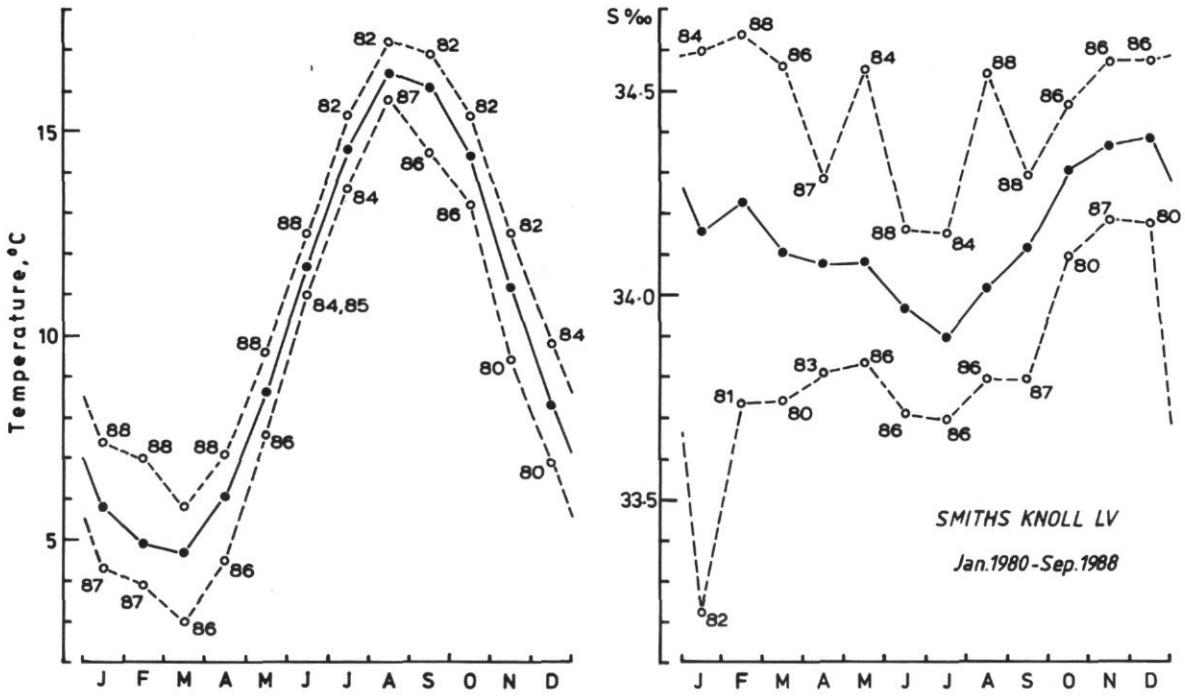

Figure 17. Monthly surface temperature and salinity means and extremes at the Smiths Knoll lightvessel (52°43′N 2°18′W), January 1980–October 1988.

Sea are discussed in a number of papers in this volume (e.g. Svendsen and Magnusson, Dickson, Colebrook and Svendsen, and the CPR Survey Team). In the later years of the decade Dickson *et al.* show that anomalous easterly winds spread fresh surface water westwards from the Skagerrak and Norwegian coast to a greater extent than usual, and Svendsen and Magnusson examine the inflow of Atlantic water into the northern North Sea, with a minimum in 1978 from the Great Salinity Anomaly and a maximum in 1990. These features are seen in Figure 18 (Anon., 1990) for positions in the western inflow and the Norwegian Trench inflow (A and B, Fig. 15). A salinity maximum occurred in 1984 at both positions unaccompanied by any distinct temperature maximum. The 1989–1990 temperature rise is shown by the volumetric/areal analysis of Svendsen and Magnusson (this volume) which demonstrates that the area north of 58°N with mean temperature below 7°C at depths greater than 50 m more than doubled from 1988 to 1989. At Utsira (Fig. 19), in the same latitude as A and B of Figure 15, but in the Norwegian Coastal Current, no signal of the "Great Salinity Anomaly" was

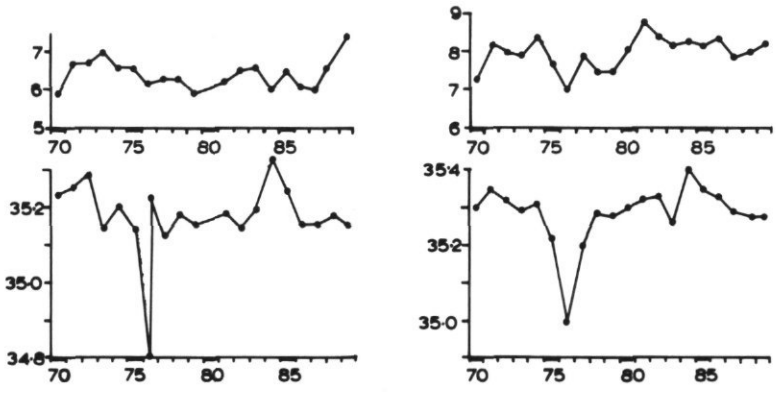

Figure 18. Summer temperature and salinity variations (1970–1989) at A (left) 59°17′N 0°22′E near the bottom, and in the core of the Atlantic inflow to the North Sea at B (right), 59°17′N 3°22′E (Anon., 1990).

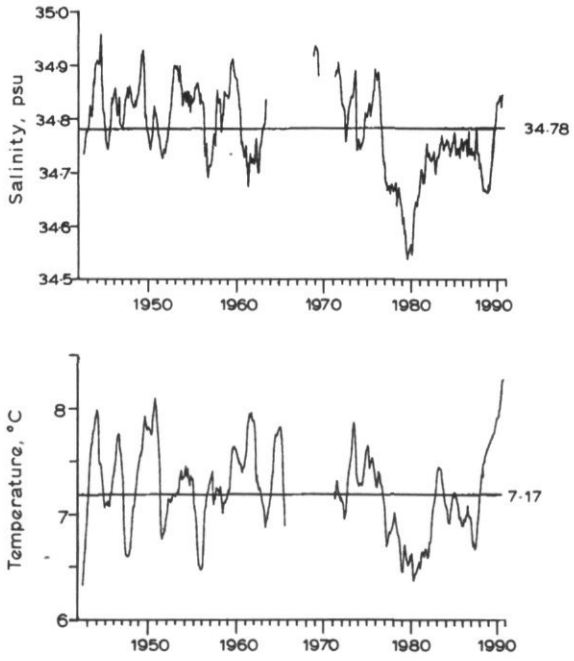

Figure 19. Surface temperature and salinity variations at Utsira (1942–1990).

seen and a longer low temperature and salinity period occurred between 1977 and 1982. Conditions remained below the 1942–1990 averages until 1989–1990, when there was a sharp increase. The delayed fluctuations in the Coastal Current may be the result of recirculation of Atlantic water from the Skagerrak.

The ICES Autumn Circulation Experiment (ACE) of 1987 found an Atlantic inflow to the northern North Sea close to the east coast of Shetland which appears to have been absent during the 1970s (Turrell, this volume). The Fair Isle Current was enhanced, and a semi-permanent eddy was established over the Fladen Ground.

Two other North Sea papers may be noted here. Becker *et al.* (1984) have compiled monthly mean temperature and anomaly charts for the previous decade which may be useful for comparison, and Koslowski (1989) provides annual figures for the number of days with ice on the German coast. In the winters of 1981–1982 and 1984–1985 to 1986–1987 the number of days with ice was above average, but in 1982–1983, 1987–1988 and 1988–1989 ice was minimal.

Norwegian and Greenland seas

In the southern Norwegian Sea a major topic of the later 1980s has been the ICES *N*orth *A*tla*n*tic *S*ea *E*xcha*n*ge (NANSEN) Project. Data were collected by a large number of participants between 1986 and 1989, and early findings are available as 1991 ICES Statutory Meeting papers. Results from investigations of overflow into the Atlantic under this programme have been noted above.

Between 63° and 64°N in the Norwegian Atlantic Current, summer standard section data since 1978 (Loeng *et al.*, this volume) show a gradual increase in temperature and salinity to a maximum in 1983, followed by a colder and fresher period with a minimum in 1988 and rising values thereafter. Spring data (mainly from April) had similar increases until 1983, but although subsequent variations had similar features to the summer data, they were less pronounced.

Off Lofoten and in the West Spitsbergen Current at 76°20′N summer temperature data also had similar features, with maxima in 1983–1984 and at the end of the period. The relative warming during 1989–1990 increased northwards. Although there was a clear timing delay in the 1970s salinity anomaly between the Faroe–Shetland and the Svalbard–Barents Sea areas (Dickson and Blindheim, 1984), no similar delay is evident for the later variations. As with the southern sections, after a clear increase from the minimum, salinity changes were less prominent but there were small contemporary maxima.

Long-term temperature and salinity observations from Ocean Weather Station "Mike" (66°N, 2°E, see Fig. 15) continued during the decade and are reported in this volume by Gammelsrød, Østerhus and Godøy.

In North Icelandic waters 1980 and 1984–1987 were relatively warm periods, while 1981–1983 and 1988–1990 were cold and fresh. Decadal average temperature was still about 2°K lower, and salinity about 0.4 psu below, pre-1960 means (Malmberg and Kristmannsson, this volume). The flux of low-salinity water north of Iceland in relation to the wind stress curl over the Nordic Seas is discussed by Jónsson (this volume).

Intensive work carried out within the Greenland Sea Project (GSP) has followed from the ICES Deep-Water Project of 1981–1982. Fieldwork began in 1987, and an intensive phase covered a seasonal cycle from 1988–1989, covering many aspects of water mass formation, circulation, dynamics of fronts, air-sea interaction, and ice formation. The large GSP data set is now being analysed. No renewal of deep water was observed during the project period, and the warming of the deep water observed in 1982 (see Fig. 5) had continued, indicating that there had been no renewal since then (GSP Group, 1990). This relatively rapid warming of the deep water in the absence of renewal may provide a signal for transfer to neighbouring basins, and possibly the deep Atlantic, giving important information on the response rate of the deep ocean to an upper ocean warming trend.

Measurements of the draft of ice floes in the East Greenland Current have begun, using upward-looking sonars. First results indicate a mean annual draft of 2.38 m (Vinje, 1990).

The Barents Sea

The 1980s conditions in the southern Barents Sea, where the influence of the Norwegian Atlantic Current predominates, have been discussed earlier. Three standard Norwegian sections and the Russian Kola section show similar trends, with temperature and salinity maxima in 1983 and 1989–1990 (Loeng *et al.*, this volume). Rich year classes of cod, haddock, and herring were recruited in 1983 with the abrupt temperature rise after the Great Salinity Anomaly (Saetersdal and Loeng, 1987). In their early years, these year classes introduced a high predation pressure upon all stages of the capelin stock, which collapsed during the mid-1980s. Cod in particular suffered from food shortage.

The end-of-decade warm period, with the highest temperatures for many years, has brought improved growth conditions, particularly for capelin. Fast growth of the large 1987 and 1988 year classes quickly restored the stock during 1989 and 1990.

Conclusions

On the whole, the 1980s appear to have been a fairly "normal" decade, with no large fluctuations. The Atlantic Subpolar Gyre remained cool although mean air temperatures were some of the highest on record. In most areas, recovery from the low temperatures and salinities of the "Great Salinity Anomaly" was taking place in the early 1980s. A cold period in the Northwest Atlantic in 1982–1985 and the Labrador Sea in 1982–1984 was connected with cold winters in the eastern Canadian Arctic, and its effects might be tentatively traced in minima occurring around the gyre at Flemish Cap in 1982–1984, station CHARLIE in 1986, in the Rockall Trough in 1986–1987, and in the cooler and fresher mid- to late-1980s in the Norwegian Sea. The North Sea was generally cool until the end of the decade, with minima during 1986 in the south. Greenland Sea ice was reported to have increased from 1987.

Intensive process studies during the decade have improved our knowledge of which variations are significant, and which are less so, and with the field programme of the World Ocean Circulation Experiment (WOCE) in the 1990s, will provide a good basis for detailed comparisons in future decades. Less satisfactory is the demise of many long-term series where observers are no longer present. New recording systems, particularly for salinity, are an urgent need if sophisticated programmes such as the Global Ocean Observing System (GOOS) are to be able to build upon the data of the past five to ten decades which ICES has played a very large part in acquiring.

Acknowledgments

The authors express their respect for the achievements of the five senior colleagues to whom this symposium is dedicated. We also thank Arthur Lee, Odd Sælen, and Lars Midttun for personal advice and encouragement over many years. Permission to reproduce figures has been granted as follows: Figures 1 and 8 "Nature"; Figure 7, Springer-Verlag; Figure 11, Royal Meteorological Society and Figure 12, the Controller of HM Stationery Office.

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ICES mar. Sci. Symp., 195: 32–39. 1992

Variability of convective conditions in the Greenland Sea

Jens Meincke, Steingrímur Jónsson, and James H. Swift

Meincke, J., Jónsson, S., and Swift, J. H. 1992. Variability of convective conditions in the Greenland Sea. – ICES mar. Sci. Symp., 195: 32–39.

Recent observations and data compilations show decadal and interannual variations in the depth of wintertime convection in the Greenland Sea. In a qualitative study the fluctuations are related to changes in wind and thermohaline forcing. Changes in both wind-stress curl and sea-ice cover concur with the results from hydrographic observations indicating that no renewal of deep water has taken place during the 1980s.

Jens Meincke: Institut für Meereskunde der Universität, Hamburg, Germany; Steingrímur Jónsson: The Marine Research Institute and University of Akureyri, Akureyri, Iceland; James H. Swift: Scripps Institution of Oceanography, La Jolla, California, USA.

Introduction

The process of convective renewal of waters in the Greenland Sea is a major influence on the thermohaline circulation of two systems (Aagaard *et al.*, 1985): on the one hand it is one of the sources for the formation of intermediate waters which leave the area as overflows across the Greenland–Scotland ridge and drive the global circulation system of the North Atlantic Deep Water. On the other hand it contributes significantly to the deep circulation internal to the Arctic Ocean/Nordic Seas system, which is confined to the area north of the Bering Strait and north of the Greenland–Scotland ridge. The division between the two systems can be defined by the sigma – 1 level of 32.785, which is found around 1500-m depth in the Nordic Seas. This is the maximum depth of winter convection that determines whether convective renewal is limited to intermediate water or affects both intermediate and deep levels.

With this background a study on the variability of convective depths in the Greenland Sea is important for the role of the oceans in climate. This contribution will illustrate the variability and discuss the mechanism causing it.

The convective system

A much improved database and conceptual modelling have led to a fairly consistent picture of convective water mass transformation in the Arctic Ocean/Nordic Seas system. The thermohaline part is dominated by the regionally varying interaction of three members in TS-space (Aagaard *et al.*, 1985; Rudels and Quadfasel, 1991), illustrated in Figure 2: cold and fresh water from runoff and melting, cold and saline water from brine released during sea-ice formation, and warm and saline water contributed from Atlantic inflows.

In the Arctic Ocean wide shelf areas the brine from seasonal sea-ice formation accumulates in troughs and trenches (Midttun, 1985) before spilling over the shelf edge. It sinks down the slopes as dense high salinity plumes at near freezing temperatures, entraining in particularly warm intermediate water of Atlantic origin, and results in Arctic Ocean Deep Water (AODW), the warmest and most saline deep water component in the Arctic Ocean/Nordic Seas system.

In the case of the deep Greenland Sea the saline plumes from sea-ice formation enrich the relatively fresh upper Arctic layer with salt, such that subsequent plumes penetrate the halocline into warm intermediate layers. To preserve continuity, the warm water rises to the surface, where it melts the ice and thereby stops the convection. After cooling and new ice formation a new haline convection cycle can start (Rudels, 1990). This results: in a stepwise increase of convective depths (Rudels *et al.*, 1989), in extremely weak mean stratification (Fig. 2), and in Greenland Sea Deep Water (GSDW), the coldest and freshest deep water component. The other deep water components, which are TS-wise between the characteristics of those from the Arctic Ocean and the Greenland Sea, are a product of mixing between the latter (Aagaard *et al.*, 1991).

The thermohaline part of the convective system

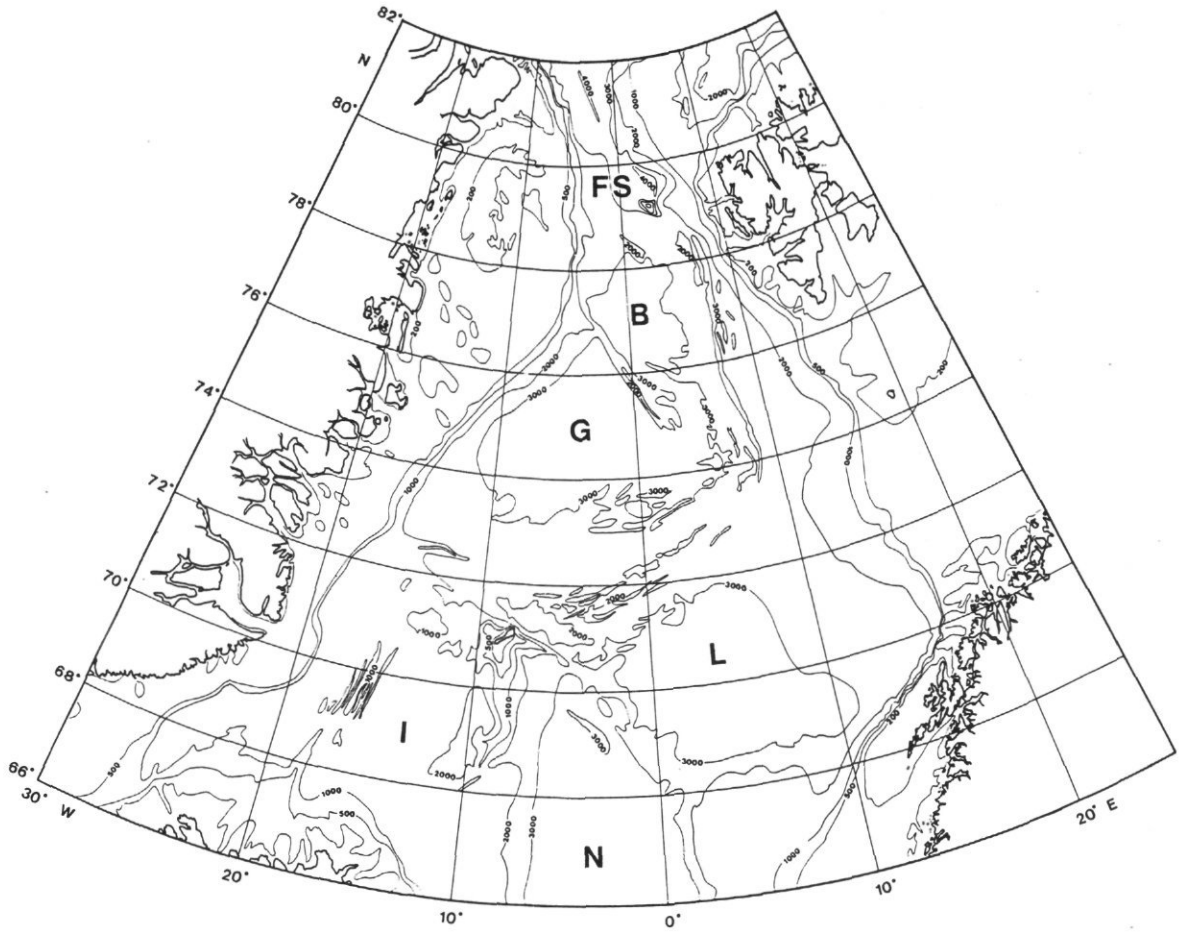

Figure 1. Topography of the Nordic Seas. FS = Fram Strait, B = Boreas Basin, G = Greenland Basin, B+G = Greenland Sea Basin, L = Lofoten Basin, N = Norwegian Basin, I = Iceland Sea Basin.

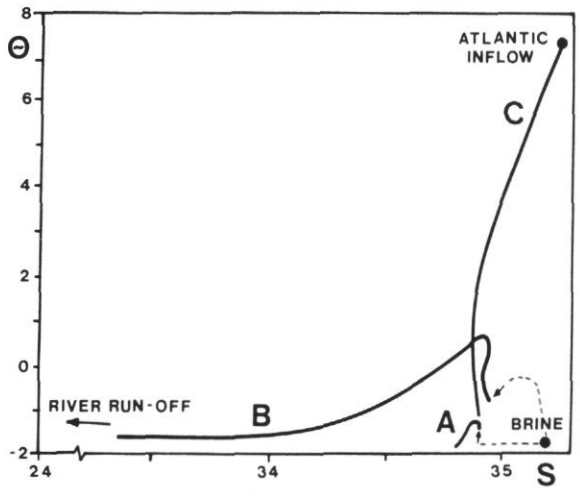

Figure 2. Schematic TS-relationships for the Arctic Ocean/ Nordic Seas system. A = Greenland Sea, B = Arctic Ocean, C = Norwegian Sea. The end members are named.

closely interacts with the wind-driven part. Next in importance to the large-scale wind forcing, which governs the flux of Atlantic waters and Polar waters throughout the system, the basin-scale wind forcing is particularly important. Legutke (1990) has shown that it is the matching between the along-isobath circulation and the contour integral of the wind stress along f/H-contours around the basis that determines the effectiveness of wind forcing. This has two consequences: with varying scales of wind forcing, different-sized cyclonic gyres are excited in accordance with the different sizes of topographic basins (see Fig. 1). For the Greenland Sea this ranges from a one-gyre situation covering both the Greenland and Boreas Basin to a two-gyre situation with separate cyclonic circulations for each of the two basins. The second consequence is related to the varying intensity of wind forcing. It results in a varying deformation of the interface between the deep and intermediate waters, i.e. a varying intensity of the "doming" of the deep water in the gyres.

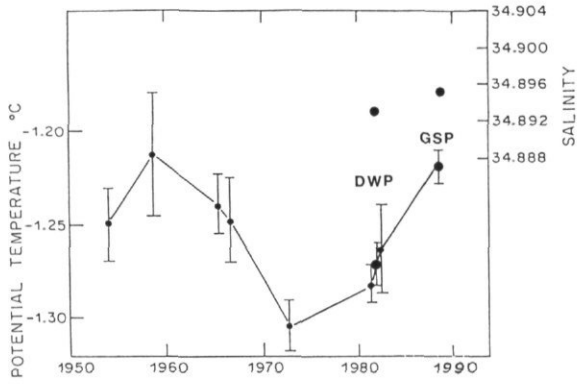

Figure 3. Time series of average potential temperature below 2000 m in the Greenland Basins. Data from stations with bottom depth exceeding 3000 m. The two single dots mark corresponding salinities obtained during the ICES Deep Water Project (DWP) (Clarke *et al.*, 1990) and during the Greenland Sea Project (GSP) (GSP Group, 1991). Θ- and S-scales were chosen such that in the event of constant deep-water density Θ- and S-variations would show as parallel curves.

Observed variations in convective activity

The reasons for the scarcity of time-series information on the convective renewal of water masses in the Arctic Ocean/Nordic Seas system are obvious: extreme difficulty of access to the Arctic Ocean and signal amplitudes close to the measurement resolution for the traditional oceanographic parameters are two major reasons. It is only since the mid-1970s that salinities have been routinely measured to 0.002 psu and that anthropogenic tracers have become available as adequate parameters. Consequently, a time series on convective renewal over a decadal time scale can only be inferred from temperature measurements in the Greenland Basin (Fig. 3). They show two warm (late 1950s and late 1980s) and one cold (early 1970s) period for the deep water, suggesting a time scale in the order of 30 years. Since the water masses of the gyre of the Greenland Basin are topographically trapped and have horizontal temperature gradients much smaller than the observed variations in time, the observed cooling can be interpreted as being dominated by renewal of GSDW, whereas the warming periods are dominated by mixing with waters around the gyre.

There are two salinity values of sufficient quality added to Figure 3. The salinity scale was chosen such that in the case of constant density the temperature and salinity curves would be parallel. This is not the case, indicating that non-isopycnic mixing occurs (Aagaard *et al.*, 1991).

The interpretation of the warming period for the 1980s in the deep water can be substantiated from increased observational activity, resulting in more detailed estimates of the annual depth of convective overturning. The Deep Water Project (Clarke *et al.*, 1990) surveyed the Greenland Sea in winter and early summer 1982 and found the convection depth at 500 m. For the period 1986–1989 (GSP-Group, 1990) convective depths were determined to be 200, 1300 and 1600 m for the Greenland Sea gyre. In combination with analyses of anthropogenic tracers covering the period 1972 to 1989, which show a cessation of deep water formation from 1980 onwards (Schlosser *et al.*, 1991), it can be concluded that the 1980s have to be characterized by no detectable renewal of waters in the Greenland Sea below 1600 m. Observations by Nagurny and Popov (1985) suggest convection down to the bottom in winter 1984. This is regarded as questionable, since salinity time series presented for the GSDW by Popov (pers. comm.) show unreasonable fluctuations.

There are two pieces of observational evidence for the cooling period in the early 1970s. Malmberg (1983) reports having observed convective conditions of the central Greenland Basin down to 3500 m in winter 1971. The other evidence comes from tritium determinations (Schlosser *et al.*, 1991), which suggest intensive deep water formation having occurred between the mid-1960s and the early 1970s.

Causes of convective variability

The description of the convective variability of the Greenland Sea has revealed decadal and interannual time scales. The literature offers only a few discussions on possible causes. Aagaard (1968) correlated long-term changes and interannual changes of intermediate and deep water temperatures with a winter cooling index derived from air temperature observations. The results support Nansen's (1906) concept of the overturning of cooled surface waters, neglecting any haline component in the system.

In discussing the wind causing variability in the convective intensity we take up Jónsson's (1991a) suggestion: an analysis of wind-stress curl over the Nordic Seas by means of empirical orthogonal functions. Jónsson (1991b) shows significant decadal and interannual variations of intensity and spatial scales. The lower modes, which compare in scale to the Greenland Sea, show interannual variability with high energy in the 1960s and early 1970s and low energy in the late 1970s and 1980s. Figure 4 is a relevant presentation of this scale. The higher modes compare in scale with the individual basins (Fig. 1), have little energy in the 1960s and 1970s, and reach high levels in the 1980s.

With respect to wind-driven circulation and convection a one-gyre situation is favoured for the Greenland Sea because of the dominance of the lower modes in the

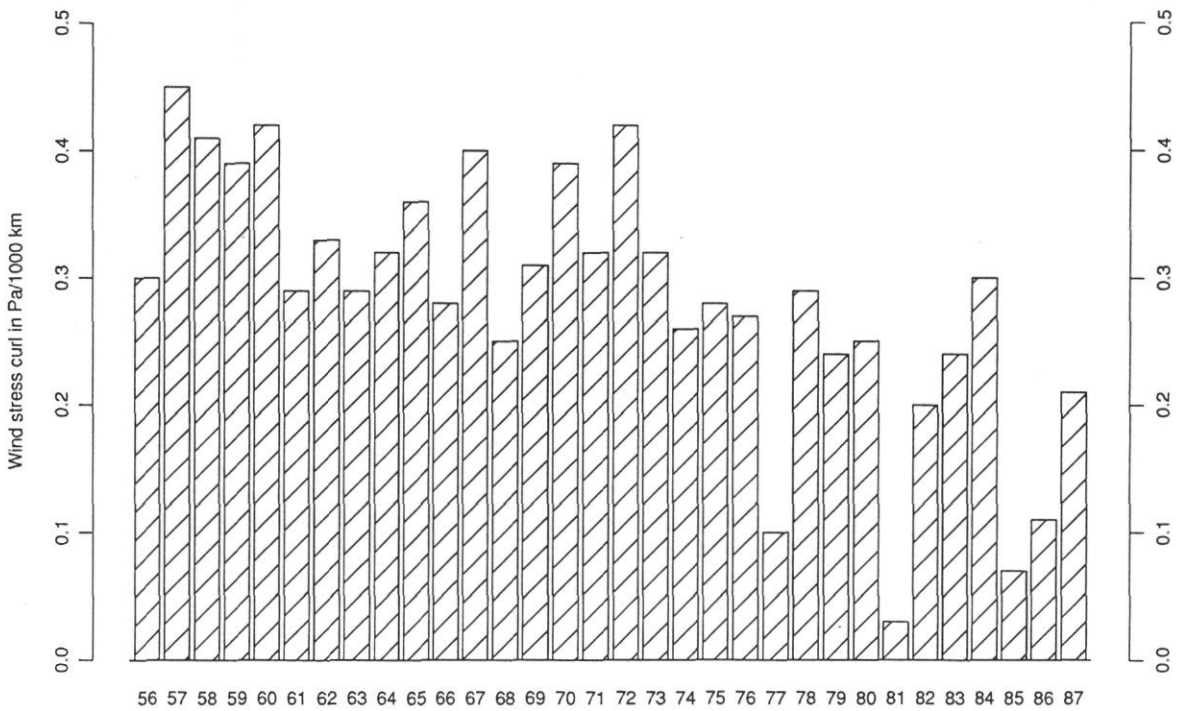

Figure 4. Interannual variations of wind stress curl over the Greenland Sea. The values are averages over an area of 200 000 km^2 in the central part of the Greenland Sea and over a one-year period from summer of year −1 to the following summer. Note the decrease in the late 1970s and 1980s.

1960s and early 1970s, with strong doming, upper layer divergence due to Ekman pumping and well developed Arctic and Polar fronts separating the convective regime from the Polar and Atlantic waters. This situation presents favourable conditions for deep reaching convection. During the 1980s, the lower mode forcing was less intense and the potential for the generation of separate gyres in the Greenland and Boreas Basin was much larger, thus repeatedly interrupting the prevailing, but weakened, one-gyre situation for certain periods. Quadfasel and Meincke (1987) indeed observed a two-gyre situation in 1986. This causes advective "leakages" of the frontal zones in the area of the gyre separation, bringing fresh Polar waters into the surface layer and warm Atlantic water into the intermediate layers of the central Greenland Sea. Both effects are adverse to convection. Aagaard and Carmack (1989) have shown this for excess fresh water input into the convective gyre and Figure 5 shows this for an intrusion of warm Atlantic water. It occurred between August 1986 and March 1987, restricting the convective depth in the Greenland Basin to 200 m because too much heat had to be removed by the convective cycles during one winter.

This makes decadal and interannual changes in the wind field one of the likely causes of corresponding changes in convective depths in the Greenland Sea – quantitative proof has still to be evaluated.

Turning to thermohaline causes of convective variability, the convective processes in the Arctic Ocean and Greenland Sea, as described in section 2 above, imply net fluxes of heat and salt as follows: in the Arctic Ocean the downslope penetration of high salinity shelf water at freezing temperatures and the entrainment of warm Atlantic waters produce a net downward flux of heat and salt for the Arctic Basins. In the Greenland Sea the entrainment of fresh water in haline plumes at freezing temperatures and the penetrative character of the plumes in the intermediate warm levels result in an upward flux of heat and salt. To compensate for gains and losses in the interior of the Arctic Ocean/Nordic Seas system the two reservoirs have to couple by exchanges through the Fram Strait. Figure 6 is a schematic diagram of this concept. Since the size of the reservoirs is vastly different, the forcing of the fluxes differs and since non-linearities of the TS-relationships are involved (see Fig. 3) the coupled system will react with oscillations. No time-series data are available, but the description of the "Great Salinity Anomaly" by Dickson *et al.* (1988) hints that such coupling exists: the authors have described an anomalous high freshwater accumulation in the area north of Iceland in the late 1960s. It was believed to have originated from an increased outflow of fresh water from the Arctic Ocean. This outflow, which Aagaard and Carmack (1989) esti-

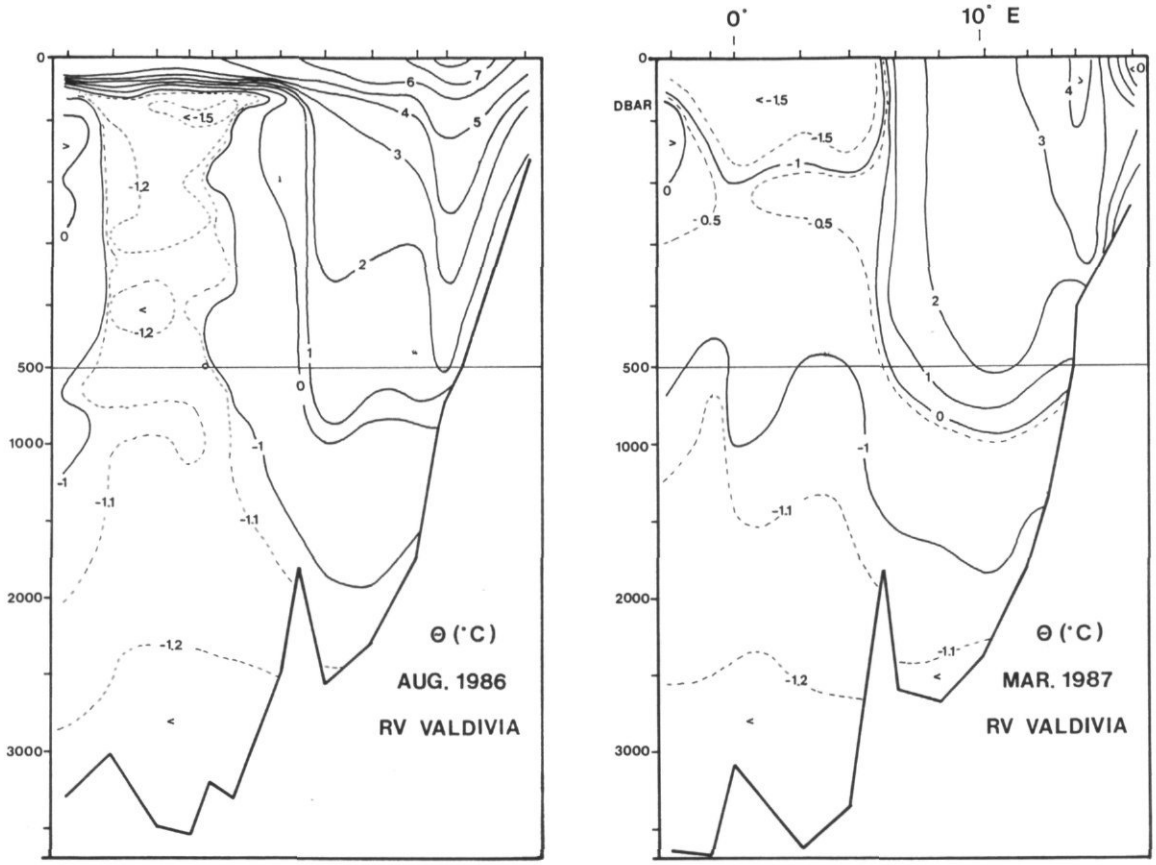

Figure 5. Temperature distribution for summer 1986 (left) and winter 1987 (right) along 74°45'N starting off at Bear Island and running due west across the central Greenland Basin. Note the warm intrusion in the interval 200 to 500 m observed in winter.

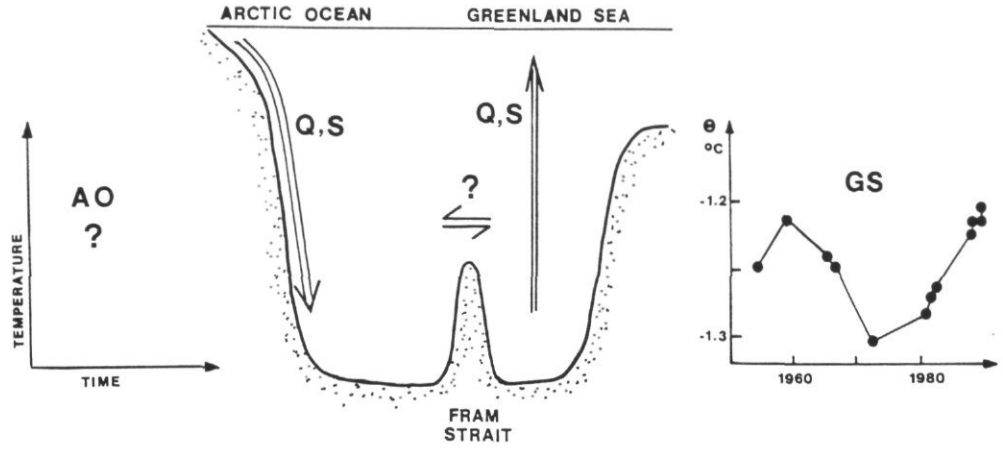

Figure 6. Schematic representation of the Arctic Ocean/Nordic Seas system as a two-basin system with different thermohaline fluxes due to different processes for deep water formation. Fluctuations are known only for the deep Greenland Sea.

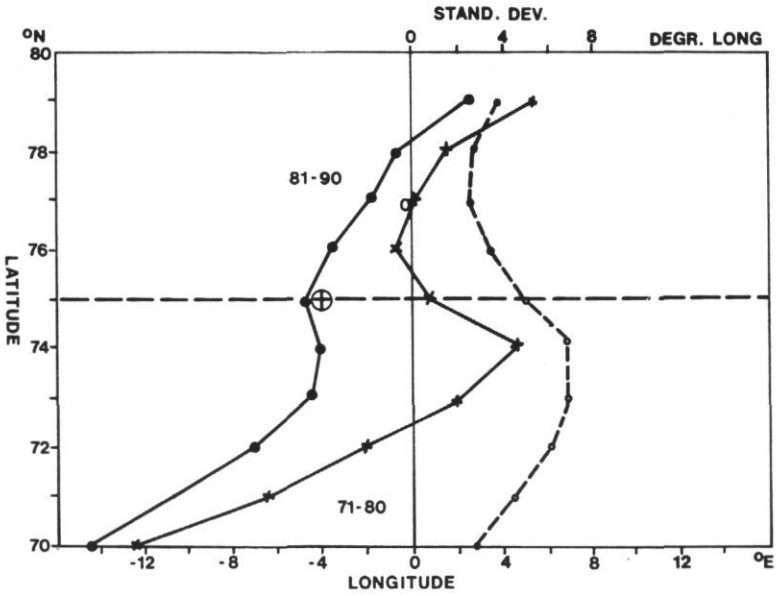

Figure 7. Extent of sea ice in the Greenland Sea. Solid lines: Wintertime (Jan to Mar) location of ice edge for the periods 1971–1980 and 1981–1990. Data sources: 1971–1980 Vinje (1984), 1981–1990 Joint Ice Center. Dashed line: Standard deviation as obtained for the period 1981–1990. No compatible estimates were possible for the 1970s. The centre for convection in the Greenland Sea is marked at 75°N, 04°W.

mated to represent a 30% increase in the mean freshwater flux from the Arctic Ocean, obviously did not immediately affect the favourable convective conditions in the Greenland Sea while passing en route to the Iceland Sea. However, while propagating from the Iceland Sea around the subpolar gyre the freshwater excess reduced convection in the Labrador Sea in 1972 and, after returning to the Greenland Sea in 1981, convective conditions became less favourable. A similar event occurred at the beginning of this century. We believe that this is some indication of the validity of the model on the generations of variability in deep water properties by thermohaline coupling of Arctic Ocean/Nordic Seas reservoirs.

The third reason for the variability is more straightforward and relates to local thermohaline forcing by sea-ice formation. As described in section 2, cycles of ice formation and melting are the surface expression of convective activity in the Greenland Sea. According to Rudels (1990) these cycles occur on time scales of the order of several days and on spatial scales of the order of kilometres. There are no routine data allowing direct observations of such cycles. Indirect indication can be obtained from the routine ice charts, which have a time resolution of the same order as can be expected for the ice cycles and a spatial resolution which is much larger. The most consistent parameter in these charts over the period 1970 to 1990 is the ice edge location. In Figures 7 and 8 the ice edge and its variance are plotted for decadal and interannual time scales.

By assuming the centre of convective activity to be at 75°N, 04°W (GSP-Group, 1990), by further assuming that a significant portion of the ice observed in the central Greenland Sea during winter has been locally formed and by implying the observed variance in the central Greenland Sea to be significantly related to ice formation/ice melting cycles, the figures can be discussed as follows: on the decadal time scale (Fig. 7) there was more ice formed during the 1970s than during the 1980s. Taking the variance of the 1980s to hold also for the 1970s, then the centre of convection was occasionally ice-free during the winter periods of the 1970s, but rather frequently during the 1980s. This result is consistent with the observed convective activity described in section 3, if more ice and more occasional open water mean more salt brine release into the system.

On the interannual time scale (Fig. 8) we find the central winter months in 1988 and 1989 just as has been discussed above for the 1980s. In 1987, however, there was so much ice that the centre of convection was permanently under an ice cover. This corresponds well with Figure 5, where the heat content of a warm intrusion of Atlantic water at intermediate depth could not be released to the atmosphere and consequently convective depths were limited to 200 m.

Conclusions

The foregoing discussion has indicated relationships between decadal and interannual variations in the depth

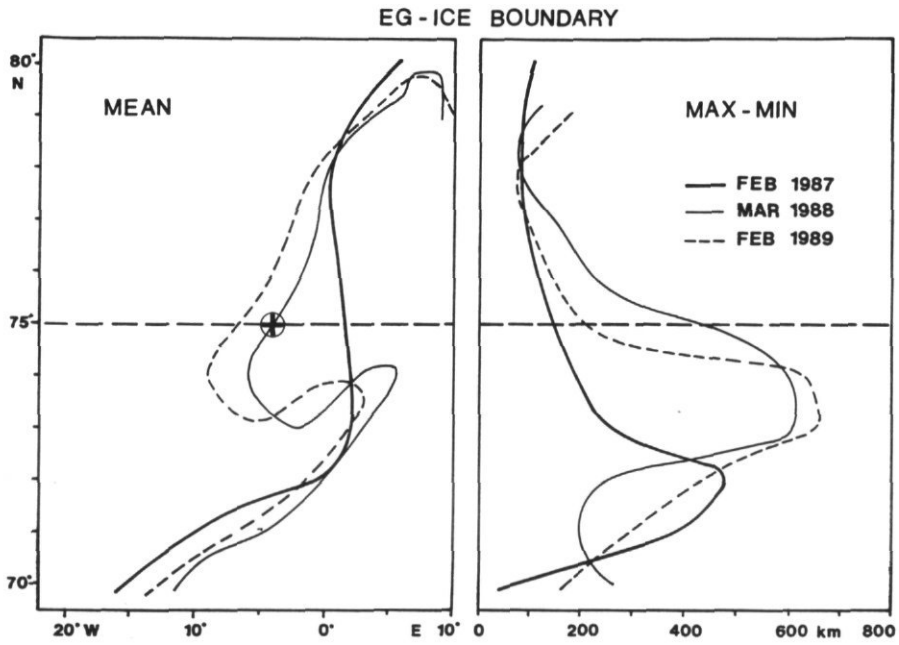

Figure 8. Monthly mean ice edge and its longitudinal variation (maximum extent minus minimum extent) for February 1987, March 1988, and February 1989. Data source: Joint Ice Center. The centre for convection in the Greenland Sea is marked at 75°N, 04°W.

of the wintertime convection and changes in forcing by wind and thermohaline processes. One cannot at present get much further along this route: the limitations on data availability and data consistency are one reason. Whereas limitations on data availability are obvious for the ocean observations, the data consistency problem concerns the longer term wind and ice data for the spatial scale of the circulation in the Greenland Sea. There have been significant changes in observation techniques for sea ice and in model-supported interpolation schemes for gridded meteorological data within the last two decades. A second reason is the fact that wind and thermohaline forcing are not independent of each other. Consequently, progress in further understanding variability in Greenland Sea convection is expected from mesoscale circulation modelling on the one hand and from continuation of relevant observations in the Greenland Sea , and hopefully in the Fram Strait and the Arctic Ocean, on the other.

Acknowledgments

This paper is based on data exchange and discussions among many colleagues working on the Arctic Ocean/ Nordic Seas system, notably K. Aagaard, E. Carmack, E. Fahrbach, D. Quadfasel, and B. Rudels.

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The Central Bank vortex in the Barents Sea: watermass transformation and circulation

Detlef Quadfasel, Bert Rudels, and Stefan Selchow

Quadfasel, D., Rudels, B., and Selchow, S. 1992. The Central Bank vortex in the Barents Sea: watermass transformation and circulation. – ICES mar. Sci. Symp., 195: 40–51.

The formation and circulation of bottom water in the western Barents Sea are examined. Freezing and induced salt ejection are identified as the mechanisms creating the dense waters. The bottom water formed above the Central Bank creates a stationary anticyclonic vortex which traps the water column above the bank and limits lateral exchanges. Only under extreme conditions, as during the strong inflow of warm Atlantic Water in the early 1980s, will the water column be completely replaced, thus inhibiting ice formation and reducing the bottom-water production. The situation described lasted until the Atlantic inflow to the Barents Sea returned to normal and the Central Bank vortex was re-established.

Detlef Quadfasel, Bert Rudels, and Stefan Selchow: Institut für Meereskunde der Universität Hamburg, Troplowitzstrasse 7, 2000 Hamburg 54, Germany.

Introduction

The Barents Sea is perhaps the most important entrance to the Arctic Ocean (Blindheim, 1989; Rudels, 1987). Atlantic and Norwegian Coastal Current waters enter between the Norwegian coast and Bear Island (Fig. 1) and continue into the Arctic Ocean, passing primarily between Novaya Zemlya and Frans Josef Land. A substantial recirculation takes place in the western Barents Sea, which returns part of the water to the Norwegian Sea (Blindheim, 1989).

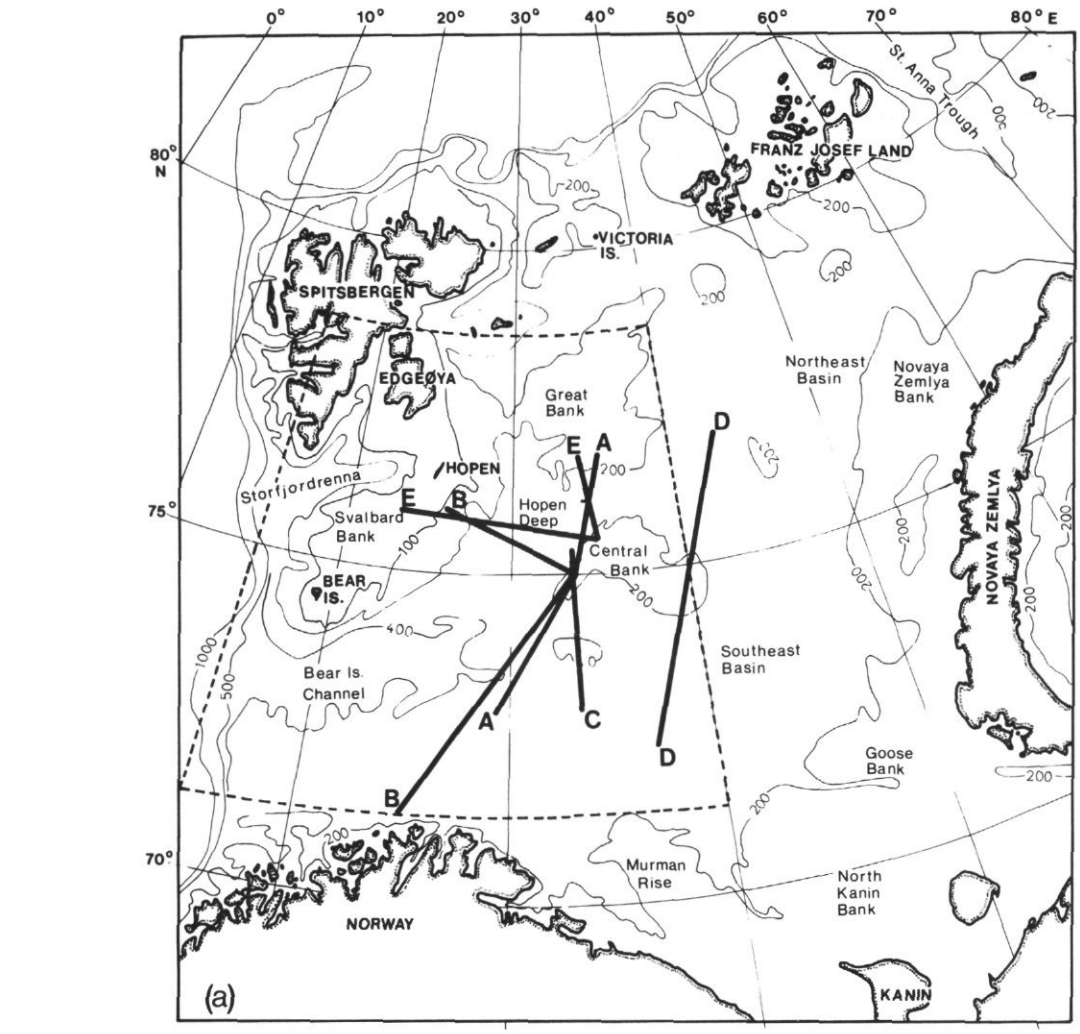
The Barents Sea is also a large shelf sea, where the inflowing water is transformed by cooling, freezing, and melting into low salinity, low density surface water and saline, cold and dense bottom waters, which are injected into the Arctic Ocean water column (Midttun, 1985; Nansen, 1906). The water mass formation takes place primarily in the shallow areas west of Novaya Zemlya and around Frans Josef Land (Midttun, 1985). Dense water is also formed in the central and western Barents Sea: over the Central Bank; on the shallow area east of Edgeöya and Hopen; and in Storfjorden (Anderson *et al.*, 1988; Midttun, 1985).

This work concentrates on dense water formation in the central and western Barents Sea. The discussion is based on CTD observations collected by RV "Lance" in 1986 on cruises arranged by the Norwegian Polar Research Institute (NP) and by the PRO MARE programme. Additional data from NP cruises with "Lance" in 1983 and 1988 as well as temperature sections taken from the Russian vessels "Pachtussov" in 1901 and "Rossiya" in 1990 have also been used. The station positions and the locations of the sections are shown in Figure 1. Because it is almost isolated from the rest of the Barents Sea, Storfjorden is not considered. Storfjorden has been described by Anderson *et al.* (1988) and Quadfasel *et al.* (1988)

The formation of dense water

The areas of dense water production can be inferred from the distribution of bottom densities (Fig. 2). Two sources are discernible: the Central Bank and the shallow area east of Edgeöya and Hopen.

Stations occupied in May 1986 east of Hopen show an almost homogeneous water column with temperatures at freezing point (Fig. 3). The slight freshening at the surface is due to spring ice melt. The cold temperatures indicate that the convection is driven by freezing and salt ejection. Saline water is accumulated on the banks, and the salinity increases throughout the winter. Cold Arctic Water dominates east of Edgeöya and only the surface summer heating has to be removed before the ice formation can begin. The water does not reach the extremely high salinities and densities found in Storfjorden (Anderson *et al.*, 1988). This could partly be due to



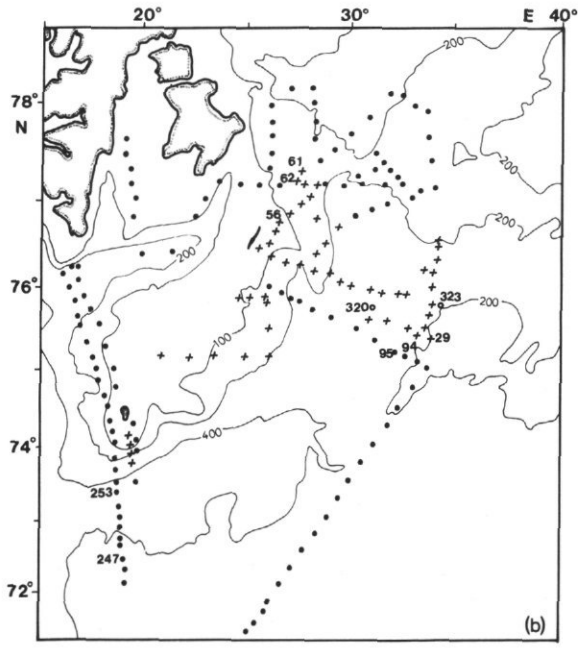


Figure 1. (a) (above) *The indicated sections are* "Lance" 1988 (A), "Lance" 1986 (B), "Pachtussov" 1901 (C), "Rossiya" 1990 (D), and "Lance" 1983 (E). (b) (left) Station positions in 1986. *Crosses represent the spring cruise, dots the summer* cruise. Numbered stations are discussed in the text. Stations 320 and 323 (open circles) were taken in 1983.

lower initial salinities, but is primarily due to strong currents and rapid exchanges of water.

In the depressions further to the northeast the vertical profiles show an increase in salinity with depth, which indicates that the dense water has been supplied from the surrounding shallower areas (Fig. 4). The temperature, salinity, and density near the bottom show large variations, suggesting also a second source area (Fig. 4).

This is most likely the Central Bank, where the salinities and densities are higher, albeit not as high as those in Storfjorden (Figs. 2 and 3). However, at no station, available to us, is a homogeneous water column to be seen. The profiles show low salinity surface water, a warmer intermediate layer of Atlantic origin, and bottom waters with somewhat lower salinities and tem-

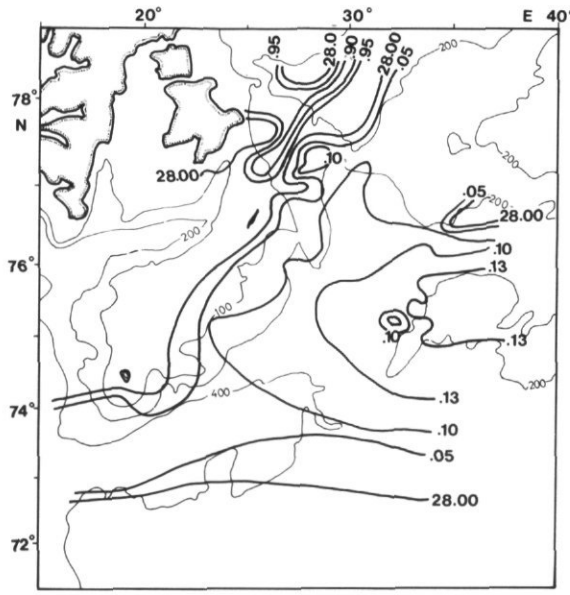


Figure 2. Distribution of bottom densities in 1986, based on all available stations.

peratures close to, but clearly above freezing point (Fig. 3).

The cold temperature of the bottom water indicates that also here ice formation is driving the convection. The fact that no station with remnants of a homogeneous water column is found implies that the convection to the bottom might be localized to shallow, unsampled areas. Another possibility is that the convection is penetrative. The removal of fresh water during ice formation increases the salinity of the upper layer and the saline water at the surface may attain such a large density anomaly that it can sink through the intermediate layers towards the bottom (Rudels, 1990). The intermediate layer is displaced towards the surface, the freezing stops and the ice begins to melt.

In winter 1986 ice formed over the Central Bank between 10 and 18 February (Norwegian Meteorological Institute ice charts). The ice cover expanded southward 150 km in one week. During this period the winds were moderate, excluding the possibility of older ice drifting over the Central Bank from the north (Fig. 5). The bank remained ice covered up to 3 March, but by 10 March the ice edge was found 150 km to the north. At the beginning of March the winds were easterly and the temperatures were extremely low (−35°C) (Fig. 5). This supports the view that the ice is removed by melting caused by warm water being brought to the surface to replace brine enriched, dense, convecting surface water.

The temperature of the bottom water will, because of entrained intermediate water be above freezing (Fig. 3). At some stations the differences are only a tenth of a degree, showing that the entrained water must be cold. Sites must exist where the intermediate layer is of small vertical extent and/or has been cooled by successive convection events. A section taken across the bank in September 1988 shows localized domes of colder water, which are, however, not properly resolved by our station grid, suggesting that the convection occurs on scales smaller than the horizontal extension of the bank (Fig. 6).

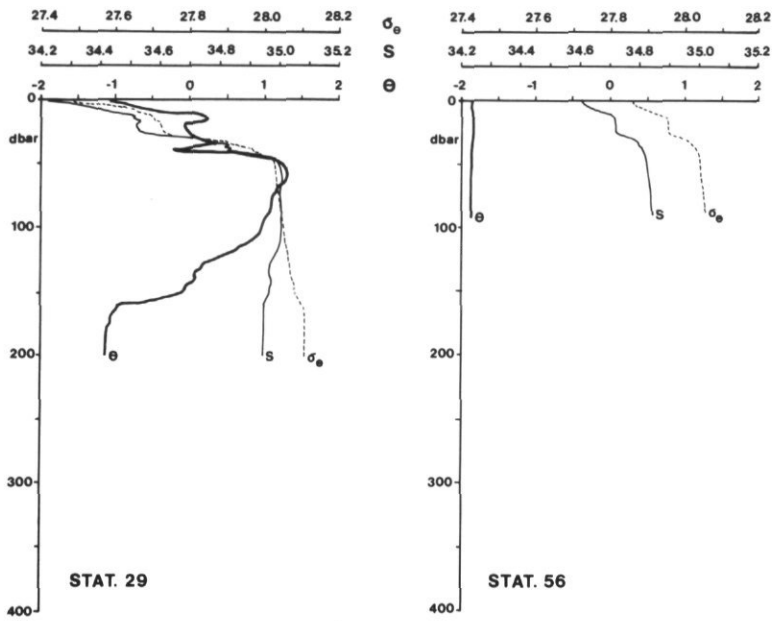


Figure 3. Potential temperature, salinity, and density profiles from east of Hopen (stn 56) and from the Central Bank (stn 29).

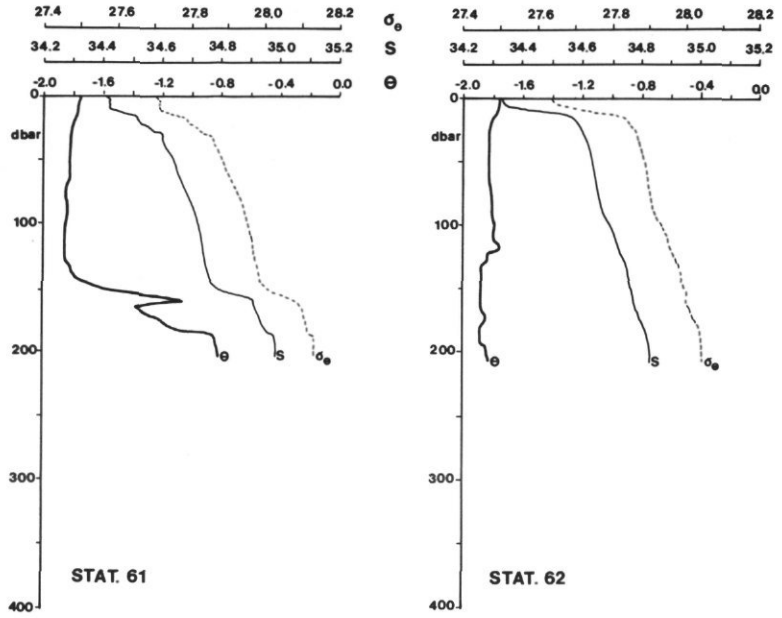
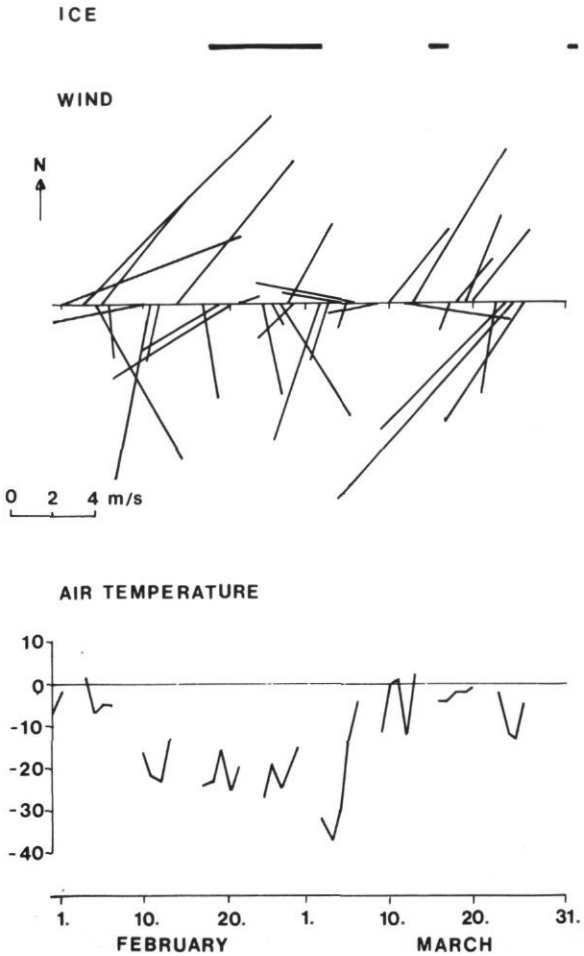


Figure 4. Temperature and salinity profiles from stations (61) and (62) east of Edgeöya.



Circulation of the bottom water

As the water sinks off the Central Bank, it becomes deflected by the Earth's rotation and an anticyclonic vortex is created. The vortex remains almost stationary, as a Taylor column, over the bank and it is permanent enough to divide the Atlantic inflow into two branches. The southern Atlantic branch flows seaward of the Norwegian Coastal Current towards the east, while the other branch turns northward to split a second time. One part follows the anticyclonal flow eastward north of the Central Bank and the other part is deflected by the Persey Current and recirculates in the Hopen Deep (Figs. 6 and 7; Tantsiura, 1959; Midttun, 1989).

The Central Bank supplies, laterally, the bottom water of the Hopen Deep, where the local thermal convection is limited to the upper 200 m. The salinity and temperature distributions (Figs. 6 and 7) indicate that the bottom water follows the rim of the Hopen Deep and exits into the Norwegian Sea south of Björnöya (Blindheim, 1989). During its transit, the bottom water mixes with incoming Atlantic Water and becomes warmer and saltier, as can be seen by comparing TS curves from the Central Bank with those from the inflow and outflow areas in the Bear Island Channel (Fig. 8). Not all dense water formed over the Central Bank returns to the Norwegian Sea. A part follows the anticyclonic flow into the eastern Barents Sea and may

Figure 5. The wind vectors and the air temperatures measured at Hopen Radio Station in February and March 1986, taken from the German Weather Service. The time when the Central Bank was covered by ice is indicated.

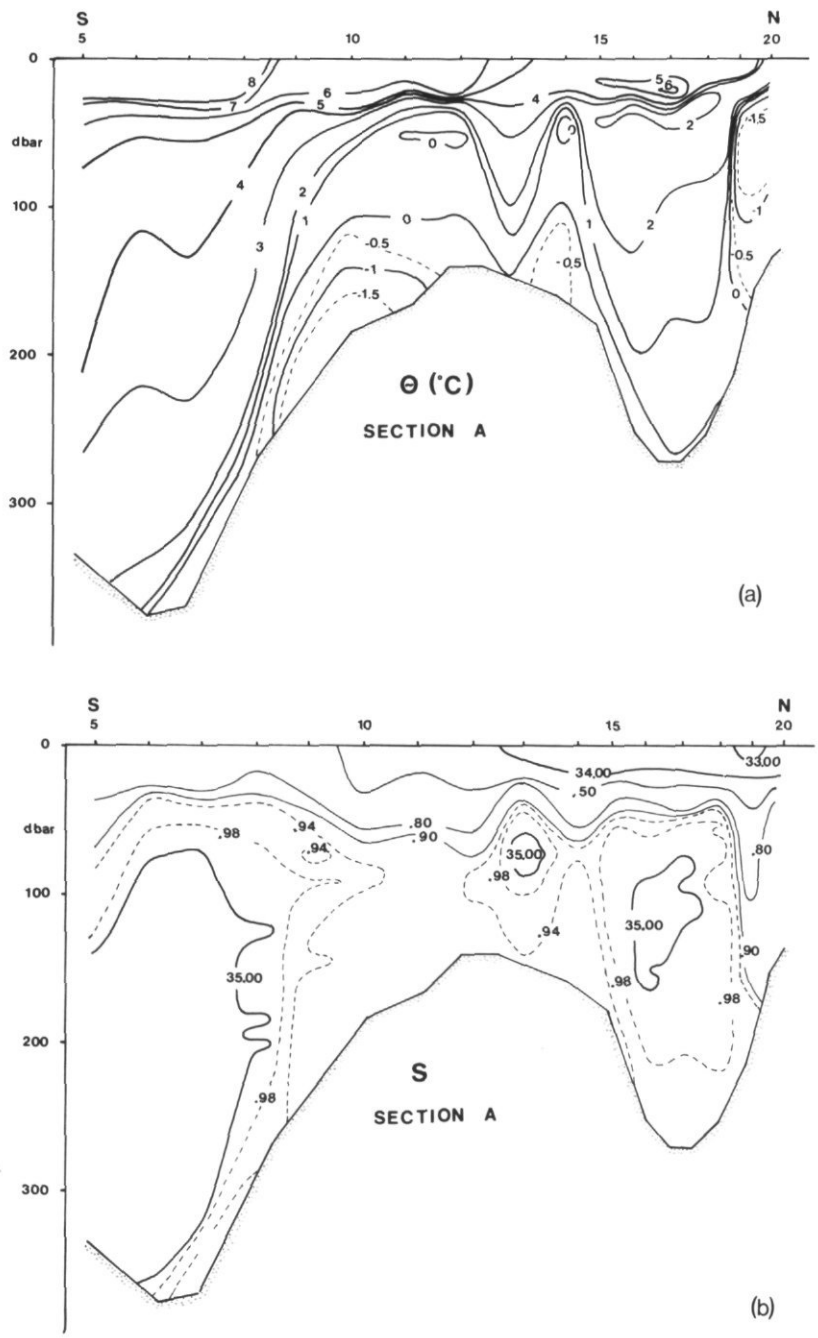


Figure 6. Vertical sections of temperature (a) and salinity (b) over the Central Bank to the Great Bank. "Lance", September 1988. Section A in Figure 1.

continue to the Arctic Ocean (Fig. 6). Some intermediate water also crosses the sill between Edgeöya and the Great Bank to form bottom water of the northern Barents Sea (Fig. 2 and also Pfirman (1984)).

The water from the western source adds, as its density level, to the outflow into the Norwegian Sea (Blindheim, 1989). Most of this water is colder, less dense, and found higher up on the slope of the Svalbard Bank (Figs. 6 and 7).

Permanence of the Central Bank vortex

The cold, dense water above the Central Bank is a persistent feature. Figure 9 shows two temperature

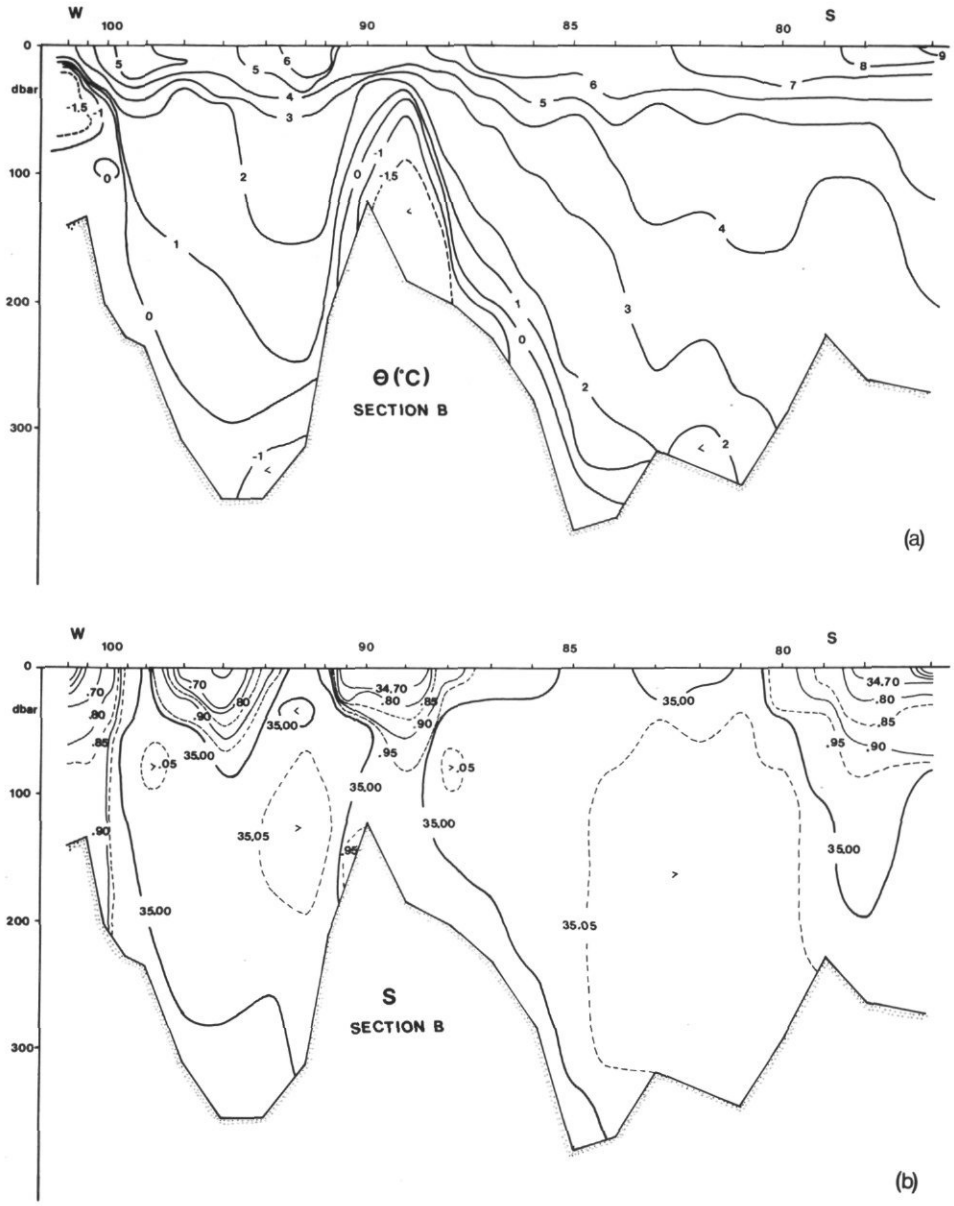


Figure 7. Vertical sections of temperature (a) and salinity (b) from the Norwegian coast over the Central Bank to the Svalbard Bank. "Lance", July 1986. Section B in Figure 1.

sections: the first, taken from RV "Pachtussov" in July 1901, the second from AI "Rossiya" in August 1990. On both sections the dome of cold water is present. It appears that the water leaking out at the bottom is replaced not through mid-level exchanges with Atlantic Water but by low salinity Arctic Water from across the Polar Front. This maintains the stratification and permits ice formation, haline convection, and reinforces the vortex.

Rapid complete renewals of the Central Bank water column are possible. In 1982 a strong inflow of warm Atlantic Water was recorded (Loeng, 1983; Midttun, 1989). Little or no ice formed over the bank during the winter 1983 (Fig. 10 and Midttun, 1989) and in the following summer neither a low salinity surface layer, nor cold, dense bottom water was observed over the bank (Fig. 11). The profiles and the TS curves observed at the bank were similar to those found in the Hopen Deep (Figs. 12 and 13). The inflow had replaced the intermediate water on the bank.

The near absence of a low salinity surface layer in the summer 1983 confirms that only limited ice formation

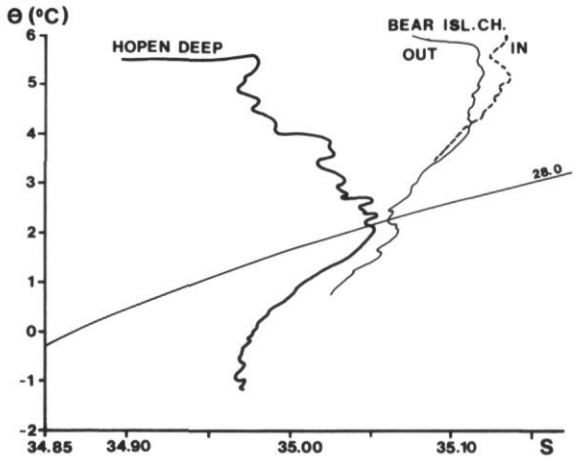


Figure 8. T–S curves from station (95) in the Hopen Deep, station (247) from the inflow area and station (253) from the outflow are in the Bear Island Channel.

and haline convection occurred before the intermediate Atlantic Water was brought to the surface. Its high temperature and the weak stratification prevented further ice formation and the deep water renewal was limited to a small volume dominated by warm entrained water (Figs. 11 and 13).

In the absence of ice production and haline convection, the anticyclonic vortex cannot be maintained. The splitting of the Atlantic inflow is weakened and the lateral, advective exchanges of the Central Bank water column can continue to bring warm water onto the bank and delay the re-establishment of the ice formation.

The warm inflow persisted at a reduced rate in 1984–1985 (Midttun, 1989). The ice edge merely touched the Central Bank in 1984 and the bank was ice-free again in 1985 (Fig. 10). Not until 1986 did the ice cover reach its long-term mean extent. The inflow of Atlantic Water was then less warm and the leaking of low salinity Arctic Water and ice across the Polar Front could compete with

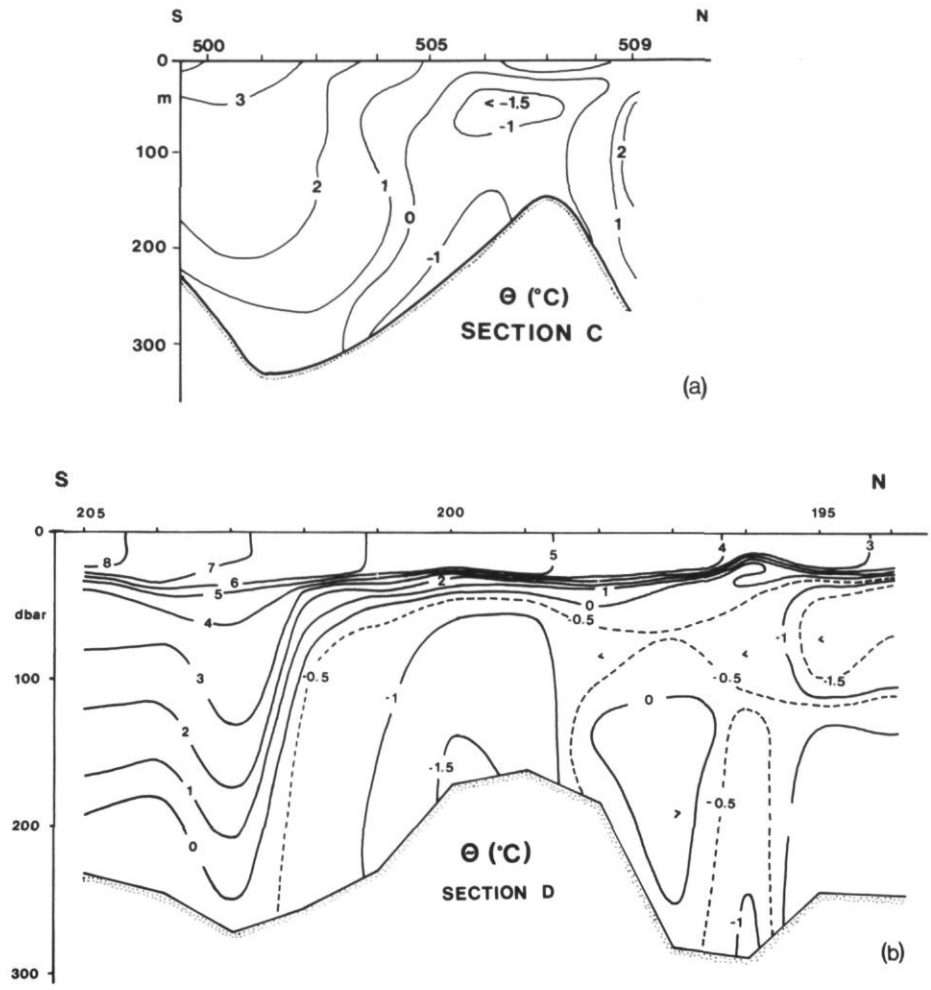


Figure 9. Temperature section taken by (a) RV "Pachtussov" in July 1901 and by (b) AI "Rossiya" in August 1990 across the Central Bank. Sections C and D in Figure 1.

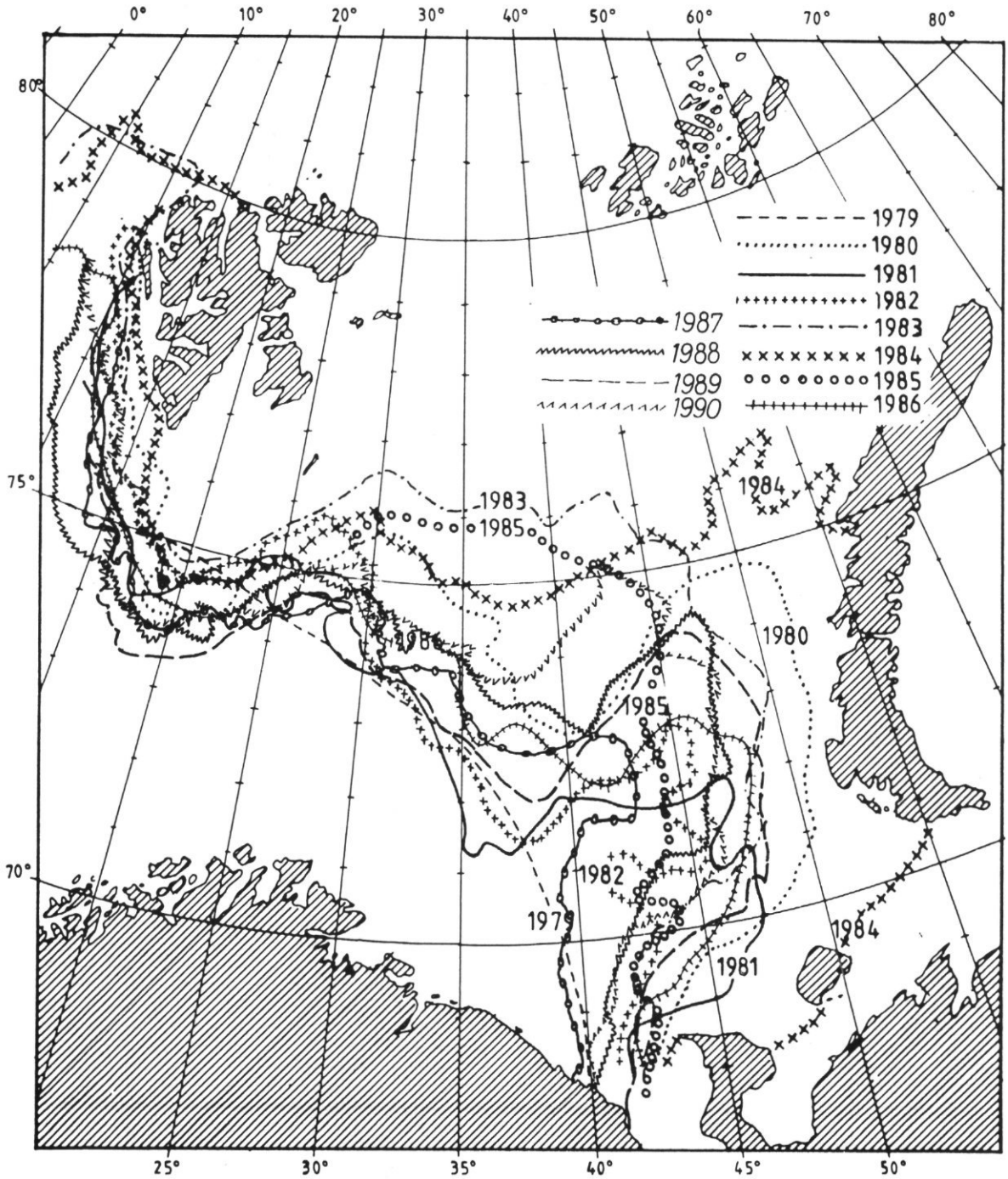


Figure 10. Maximum extent of the ice cover in the Barents Sea 1979–1990 (adapted from Midttun, 1989).

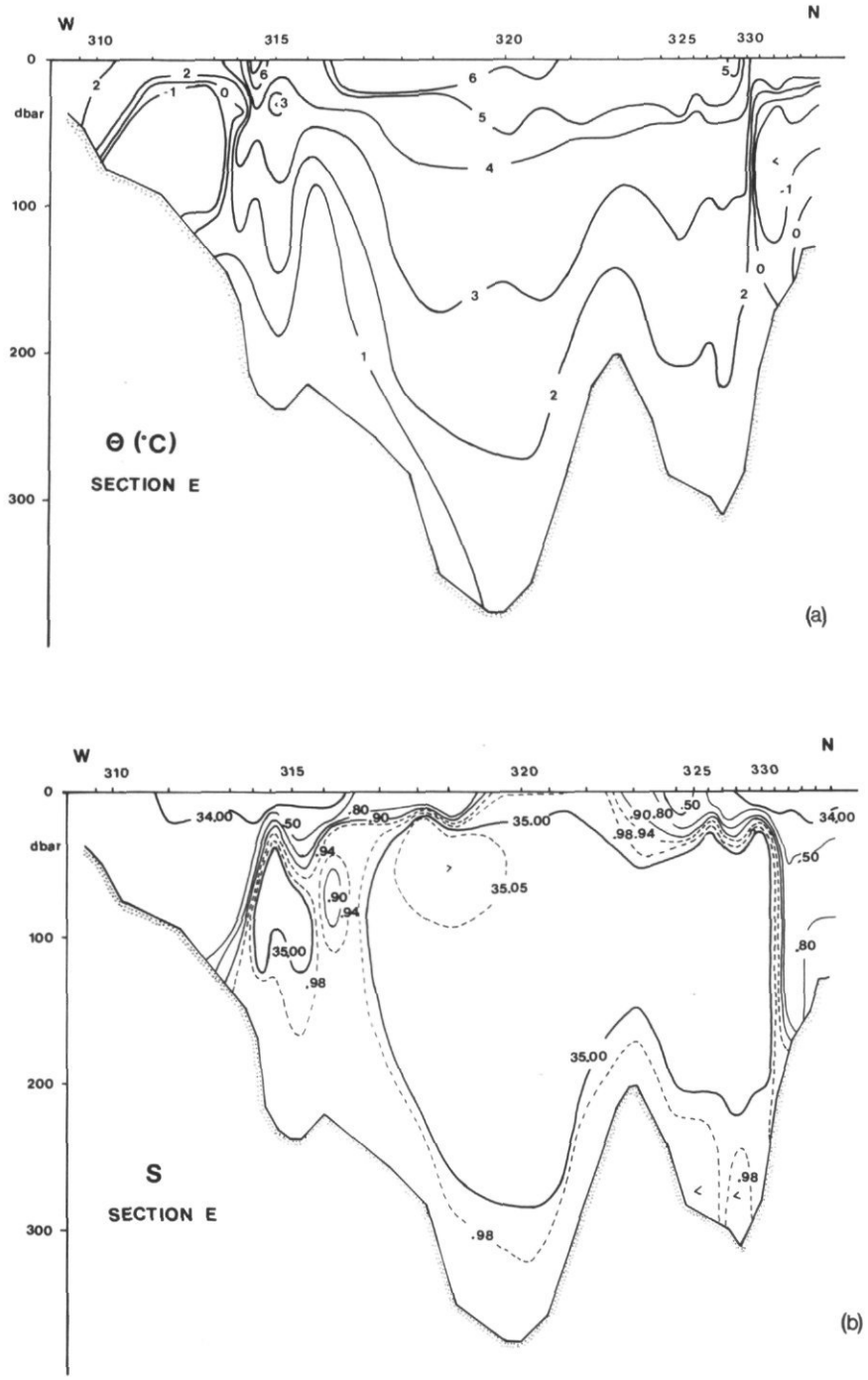


Figure 11. Vertical sections of temperature (a) and salinity (b) from the Svalbard Bank over the Central Bank to the Great Bank. "Lance", August 1983. Section E in Figure 1.

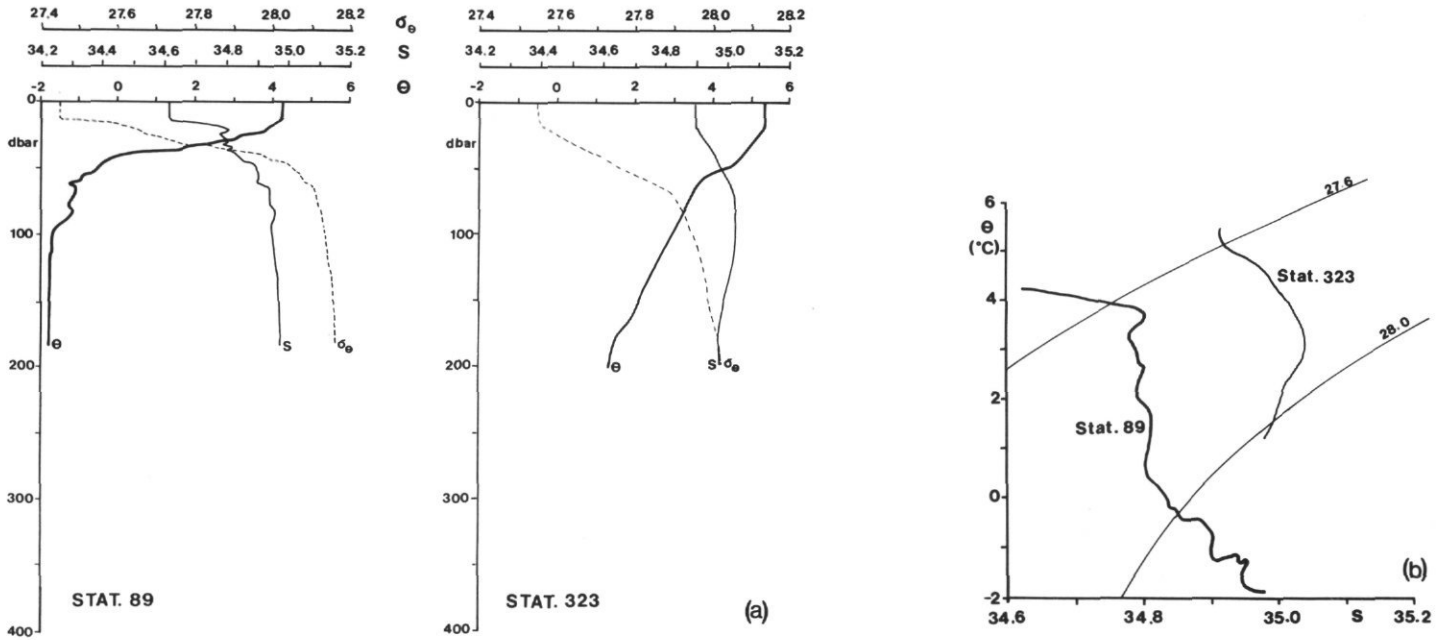


Figure 12. Temperature and salinity profiles (a) and T–S curves (b) from station (323) in 1983 and station (89) in 1986 on the Central Bank.

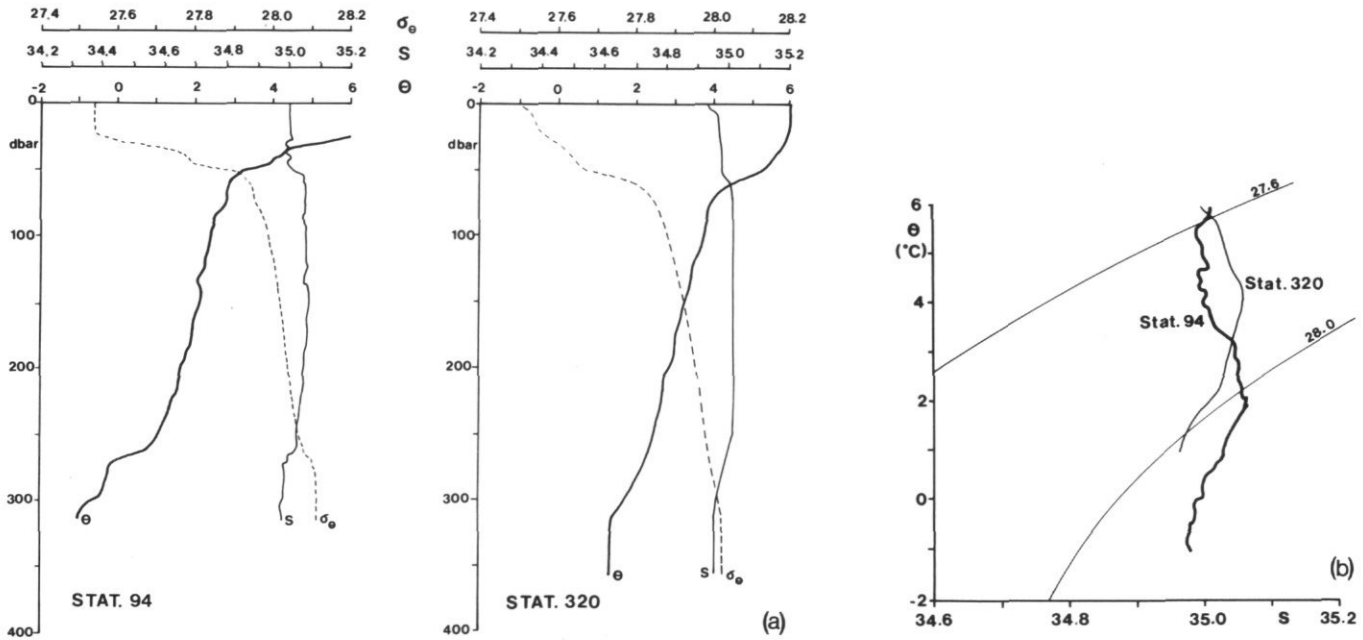


Figure 13. Temperature and salinity profiles (a) and T–S curves (b) from station (320) in 1983 and station (94) in 1986 in the Hopen Deep.

the advective transport of Atlantic Water, reform the vortex and recreate the long-term circulation pattern.

It should be kept in mind that we have no data from 1984 and 1985 or from the years before 1983. It is the correlation between ice formation and convection found above which makes us assume that the circulation and hydrography of the western Barents Sea during the ice-free years 1984–1985 resembled those of 1983, while the situation in the period before was similar to that of 1986 and 1988.

The Central Bank remained ice-free for two to three years. It is not clear if this represents the time needed to reform the stratification or whether the time span was determined by the persistence of the inflow. In view of the rapid return of the bottom water production in 1986 and of the permanence of the vortex throughout the years, it is likely that the long-term hydrographic state can be established within one season, once the disturbance is gone.

The circulation pattern in the central Barents Sea is normally stable and controlled by the Central Bank vortex. The situation which prevailed in the early 1980s was not created locally but was caused by climatic conditions outside the Barents Sea, confirming Helland-Hansen and Nansen's (1909) finding that the advection from the North Atlantic determines the convection and the climate state of the Barents Sea.

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Climatic variability in the Norwegian and Barents Seas

Harald Loeng, Johan Blindheim, Bjørn Ådlandsvik, and Geir Ottersen

Loeng, H., Blindheim, J. Ådlandsvik, B., and Ottersen, G. 1992. Climatic variability in the Norwegian and Barents Seas. – ICES mar. Sci. Symp., 195: 52–61.

Time series from several hydrographic sections in the Norwegian and Barents Seas are analysed in relation to cyclic variations and the results compared with meteorological observations. The length of the time series varies from 36 to more than 100 years. The results indicate that the time series consist of several components, with periods varying from 2.6 to 17.5 years. The 1980s were characterized by two large inflows of Atlantic water to the area, the first taking place in 1982/1983, the second in 1989/1990. Modelling results indicate that the variability in the Atlantic inflow is closely related to meteorological conditions. The warm periods are related to a stable low-pressure situation over the area with cyclonic circulation, while cold periods are linked to a high-pressure situation with anticyclonic wind circulation.

Harald Loeng, Johan Blindheim, Bjørn Ådlandsvik, and Geir Ottersen: Institute of Marine Research, PO Box 1870, Nordnes, N–5024 Bergen, Norway.

Introduction

Chavanne (1875) described the ice conditions in the Arctic back to the middle of the 16th century and made comparisons between the variability in ice coverage and current conditions. However, the first broad analysis of the hydrography, currents, and climatic variability of the Norwegian and Barents Seas were made by Helland-Hansen and Nansen (1909), who conducted comparative studies between different physical parameters and suggested that the climatic variations in these regions probably were of an advective nature. The role of advection in year to year variability has since been supported by several authors, among them Elliot (1956), Bergthorsson (1972), Dickson and Blindheim (1984), and Dickson *et al.* (1988).

Helland-Hansen and Nansen (1909) were also among the first to consider how the variability in physical conditions influence biological processes. They believed that temperature variations in the sea "are the primary cause of the great and hitherto unaccountable fluctuations in the fisheries". Investigations throughout the last 10–15 years have clearly demonstrated that there is a close relation between climatic variability and biological conditions such as recruitment, distribution, migration, and growth of commercially important fish stocks, as reviewed by Loeng (1989a). If we can understand and predict the variability in the physical conditions, it may help us to make better prognoses for the fish populations.

Only a few authors have considered the causes of this climatic variability. Again, Helland-Hansen and Nansen (1909), with their broad discussion of the physical conditions in the Norwegian and Barents Seas, were among the first to touch on this problem. In later publications (Helland-Hansen and Nansen, 1912; Nansen, 1920), they discussed the variability in atmospheric and oceanic circulation, compared air and water temperature with ice conditions, and discussed the importance of bottom-water formation. These same problems have later been discussed by, for example, Rodewald (1972), Dickson *et al.* (1988), and Mysak *et al.* (1990). However, there is still no clear answer to how meteorological and oceanographic conditions interact. During recent years, results from numerical models have been used in order to study this type of relation (Ådlandsvik, 1989b; Ådlandsvik and Loeng, 1991). Especially in the Barents Sea, processes like formation of dense bottom water may be important for the inflow activity of Atlantic water. Starting out with Helland-Hansen and Nansen's theory on deep water formation, Midttun (1985) supported this idea.

The present paper deals with the climatic variability in the Norwegian and Barents Seas, emphasizing the 1980s. Analyses of time series from different hydrographic sections and meteorological stations have been carried out in order to find cyclic variations. In addition, model results are used to show the close relationship between the atmospheric and oceanic circulation. The

causes of the variability in Atlantic inflow to the area are also touched upon.

Material and methods

In this paper we have used oceanographic and meteorological data from various hydrographic sections and meteorological stations, located as indicated in Figure 1. Sea-surface temperature is observed daily at some coastal meteorological stations. Some of these observations series were initiated by Professor H. Mohn in 1867, and referred to in Mohn (1887). Monthly means have been published by the Norwegian Meteorological Institute since 1870 (Anon., 1874–1980). For the present paper we have used monthly means of sea-surface temperature from two stations, Ona and Myken (Fig. 1). In addition, we have used monthly mean values of air temperature and atmospheric surface pressure from two meteorological stations, Bjørnøya (Bear Island) and Fruholmen.

The Institute of Marine Research, Bergen, has maintained observations on several standard sections in both the Norwegian and Barents Seas in order to see how climatic signals propagate in the current system. For some sections there exist time series back to the early 1950s. However, regular observations have only been taken once a year, in the Barents Sea in late August or early September. Since 1977, however, regular observations have been carried out more frequently, at the Fugløya–Bjørnøya section six times a year and at the Vardø–N section four times each year. In the Norwegian Sea there are only continuous regular observations from summer back to 1978. All sections used are shown in Figure 1.

The longest time series of temperature in the Barents Sea is the Russian records on the Kola section (Fig. 1), which started as early as 1900. Two periods were missing in the observation series, 1906–1919 and 1941–1944. Bochkov (1982) has filled in these gaps by a combination of interpolation and other observations. The period 1906–1919 has quarterly values, while the rest of the time series have monthly values. In our analysis we have used the monthly values from 1921 to 1990. The data from recent years have been provided by PINRO, Murmansk.

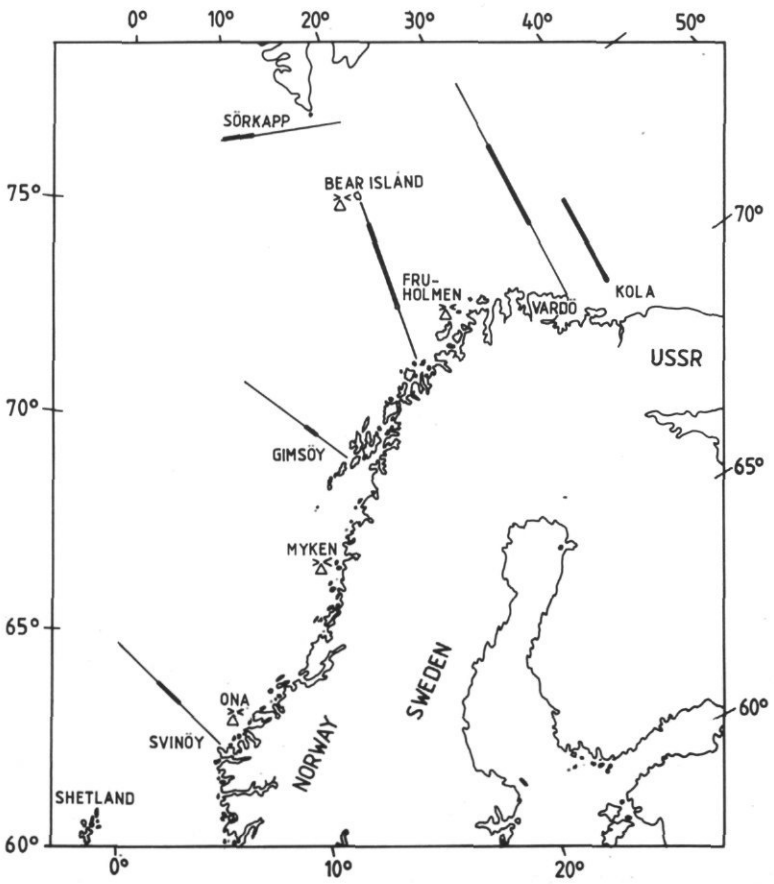

Figure 1. Location of sections and stations. The part of the sections used in the calculations is indicated by a thick line. The triangles indicate meteorological stations.

Table 1. Time series which have been analysed by means of spectral analysis. AT – air temperature, AP – air pressure, SST – sea-surface temperature, T – (mean) temperature and S – (mean) salinity.

Station/section	Parameter	Observation period	Obs y^{-1}
Bjørnøya	AT	1920–1990	12
Bjørnøya	AP	1946–1990	12
Fruholmen	AT	1954–1990	12
Fruholmen	AP	1954–1990	12
Myken	SST	1922–1986	12
Ona	SST	1868–1990	12
Kola-section	T_{0-200}	1921–1990	12
Vardø-N	T_{50-200}	1953–1990	1
Vardø-N	S_{50-200}	1953–1990	1

Means of temperature and salinity have been calculated for the main Atlantic inflow through the various sections using the method described by Blindheim and Loeng (1981). These means are prepared for the depth interval 50–200 m along the parts of the sections which are marked in Figure 1. In the Kola section we have used means from 0–200 m. The intention in presenting the time series as "box means" is to filter out small-scale spatial variability.

Spectral or Fourier analysis is used to find the principal periodicities in the time series. A series is decomposed to a sum of sine and cosine components of different amplitudes and wavelengths. The coefficients of these components are known as the discrete Fourier transforms of the series. The periodograms at the different wavelengths are worked out from the Fourier transforms and compared. This procedure is done for both the original series and the series where the strong seasonal variations have been removed by a 12 month moving average filter. The computations are done with the SAS package (SAS Institute, 1988). For working out the 12 month moving averages the X11 procedure is used, while the Fourier analysis is done with the Spectra procedure. The time series analysed are listed in Table 1.

Ådlandsvik (1989a, b) describes a numerical wind-driven model for the Barents Sea. The same model is used here to compare the variability in sea temperature with fluctuations in Atlantic inflow due to atmospheric conditions. The atmospheric forcing is taken from the "Hindcast Archive" of the Norwegian Meteorological Institute (Eide *et al.* 1985). This archive contains six-hourly values on a 75 km grid covering all seas surrounding Norway.

Results

Figure 2 shows a three-year running mean of temperature in the Kola section for the period 1900–1990. During this period the temperature has varied considerably. The main part of the period 1900–1920 was cold. The 1920s were variable, while most of the period 1930–1965 had temperatures above the long-term mean value, with a maximum in the late 1930s. Since the late 1940s, a general long-term temperature decrease has taken place. The last 25 years are characterized by large variability, which to some extent is difficult to detect in Figure 2.

Figure 3 shows more details in the variability in both temperature and salinity of the inflowing Atlantic water to the Barents Sea during the last 15 years. The cold period which started in 1977 lasted to the beginning of 1982. The temperature started to increase as early as the summer of 1981, while farther east, at the Kola section, the warming started in March/April 1982 (Fig. 7). The warm period lasted for only three years, and was followed by four years with temperatures below average. In early 1989 a new rapid temperature increase occurred in the whole southwestern Barents Sea (to about 35°E). Since then the temperature has been almost 1°C above the average. In contrast to 1982–1983 the warming in 1989 occurred simultaneously in the whole area east to the Kola meridian.

After the salinity minimum in 1978–1979 (Fig. 3), there was a fairly rapid increase in salinity up to 1980,

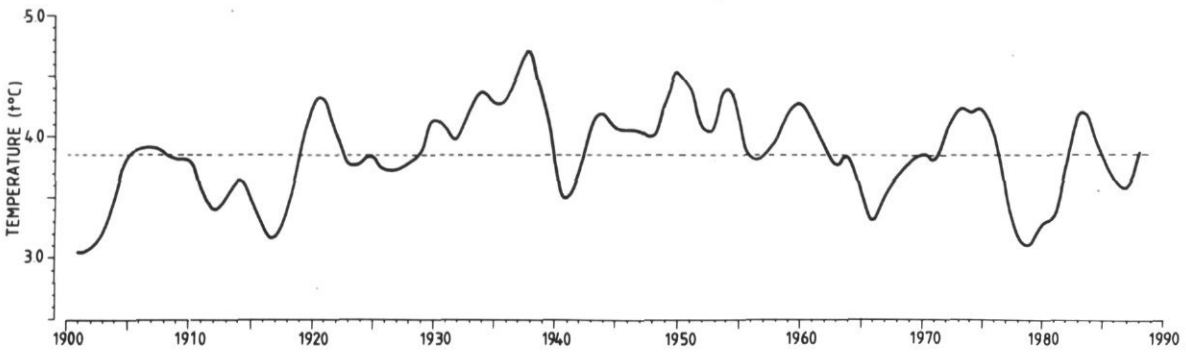

Figure 2. Three year moving average of yearly temperature in the Kola section during the period 1900–1990, based on data from Bochkov (1982) and from PINRO, Murmansk (pers. comm.). The broken line indicates the mean value for the period 1921–1980 (Loeng 1991).

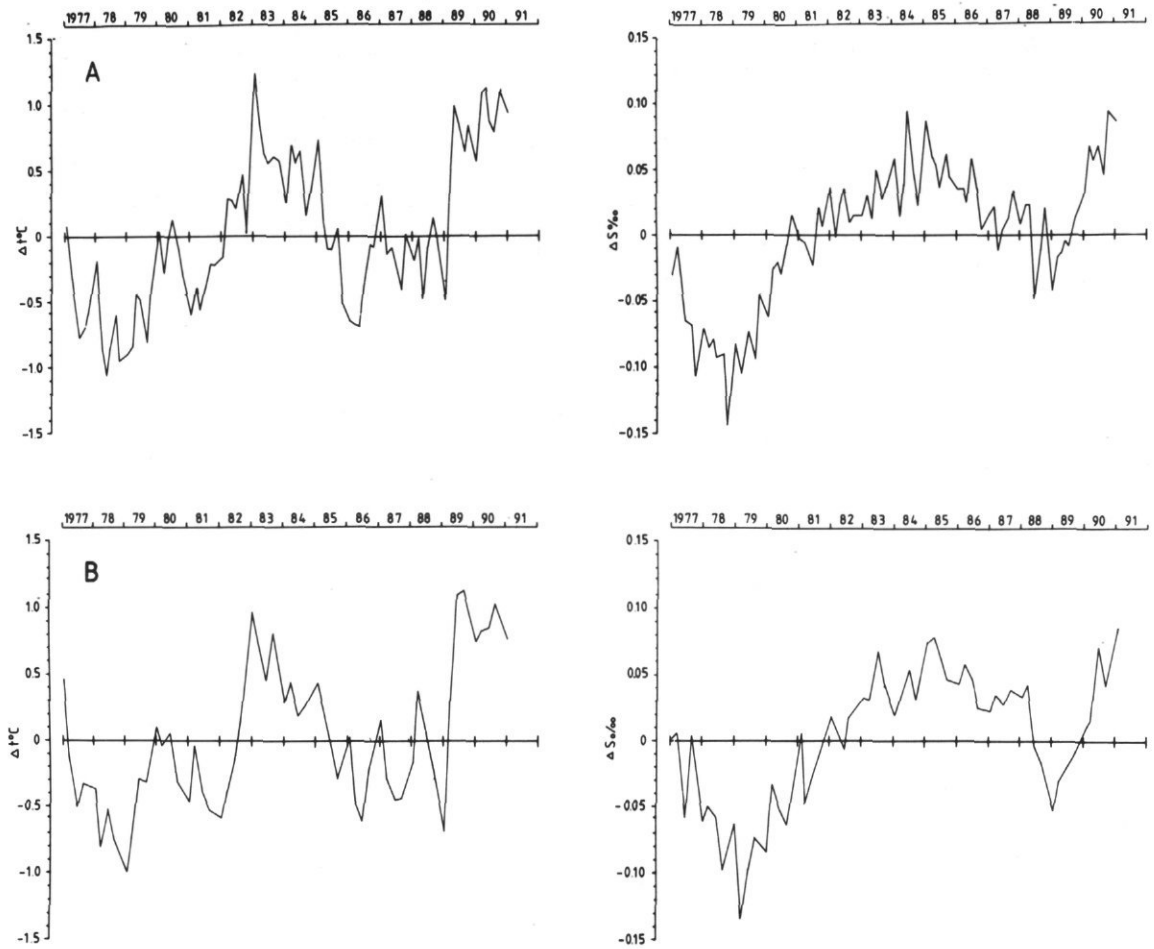

Figure 3. Observed temperature (left) and salinity anomalies (right) in the Fugløya–Bjørnøya (A) and Vardø-N (B) sections in the Barents Sea for the period 1977–1990.

followed by a much slower increase until maximum was reached in 1984. Then there was a decrease in salinity up to 1989, when a new increase in salinity started. The main features in the variability of salinity follow the temperature fluctuations, but Figure 3 also indicates small differences.

Since the sections in the Norwegian Sea have only one or two regular observations each year, it is difficult to state exactly when changes take place in these sections. Figure 4 shows variations in temperature and salinity of Atlantic water in the three sections Svinøy-NW, Gimsøy-NW, and Sørkapp-W. Summer observations in the Svinøy section showed a gradual increase in both temperature and salinity to a maximum in 1983, followed by a cooler and fresher period toward a minimum in 1988. During 1989/1990 there was again a rise toward warm and saltier conditions.

In the other two sections, summer data also showed high temperatures associated with the peak in 1983. In the Gimsøy section the main temperature increase occurred between 1982 and 1983, as in the Barents Sea. In the Sørkapp section the maximum occurred in 1983/1984, but this section also showed a separate maximum in 1980 with an almost equally high temperature as in 1983/1984. This peak in temperature was followed by a salinity maximum one year later, which showed the highest salinity of the decade in the Sørkapp section. Also, in the Gimsøy section, the 1981 maximum appeared in salinity, but it was considerably less prominent. In the Svinøy section, as well as in the Barents Sea (Bjørnøya section, Fig. 3), it was hardly traceable.

The most outstanding salinity maximum in the Svinøy section, which peaked in 1983, had a different development with a northward decrease in prominence and in the Sørkapp section it was not distinguishable. There, the general trend was a fairly gradual decrease after the peak in 1981 until the minimum in 1988. The rise in temperature and salinity during 1989 and 1990 developed similarly in all three sections, with relatively

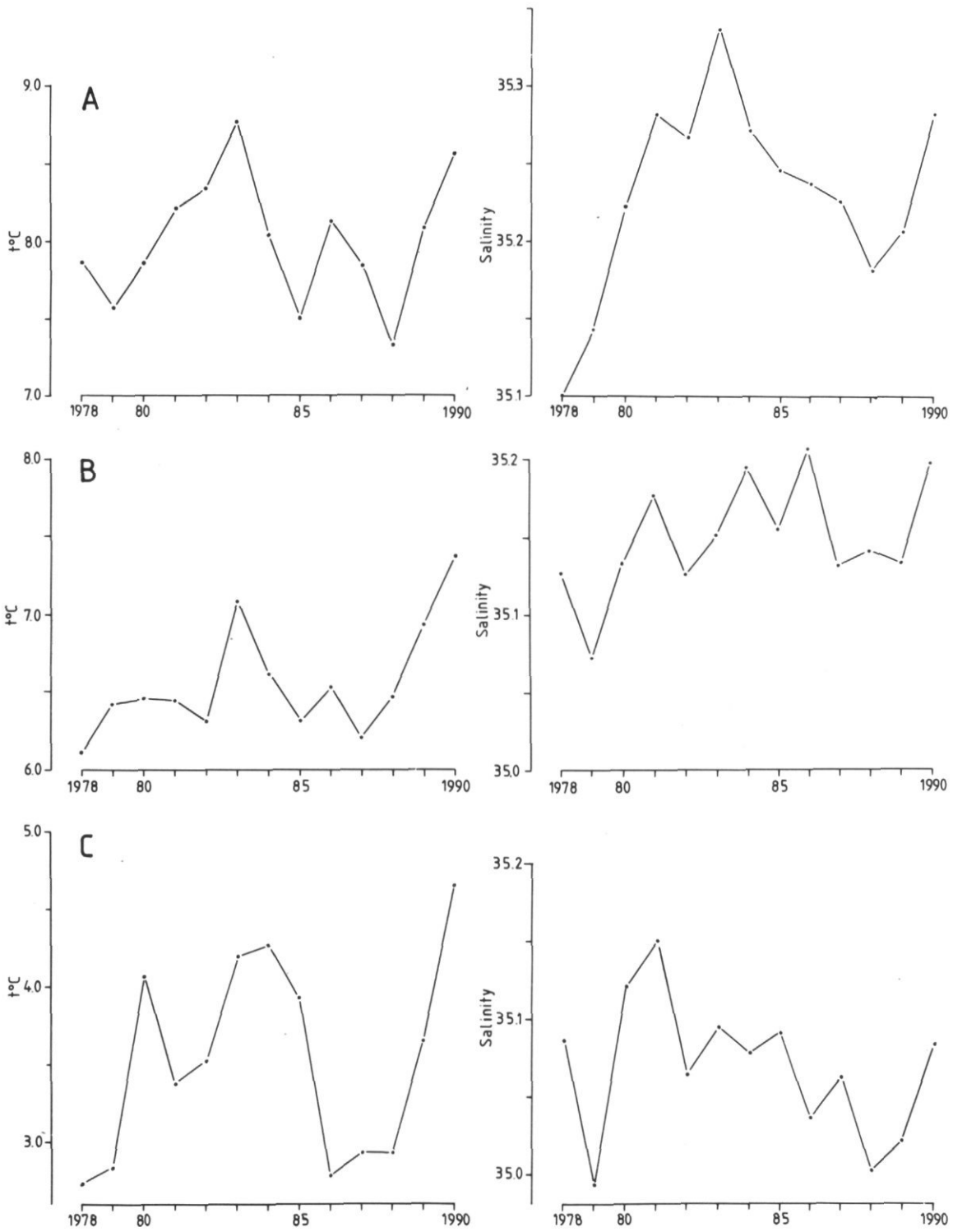

Figure 4. Temperature and salinity during summer in the three sections Svinøy-NW (A), Gimsøy-NW (B), and Sørkapp-W (C) in the Norwegian Sea in the period 1978–1990.

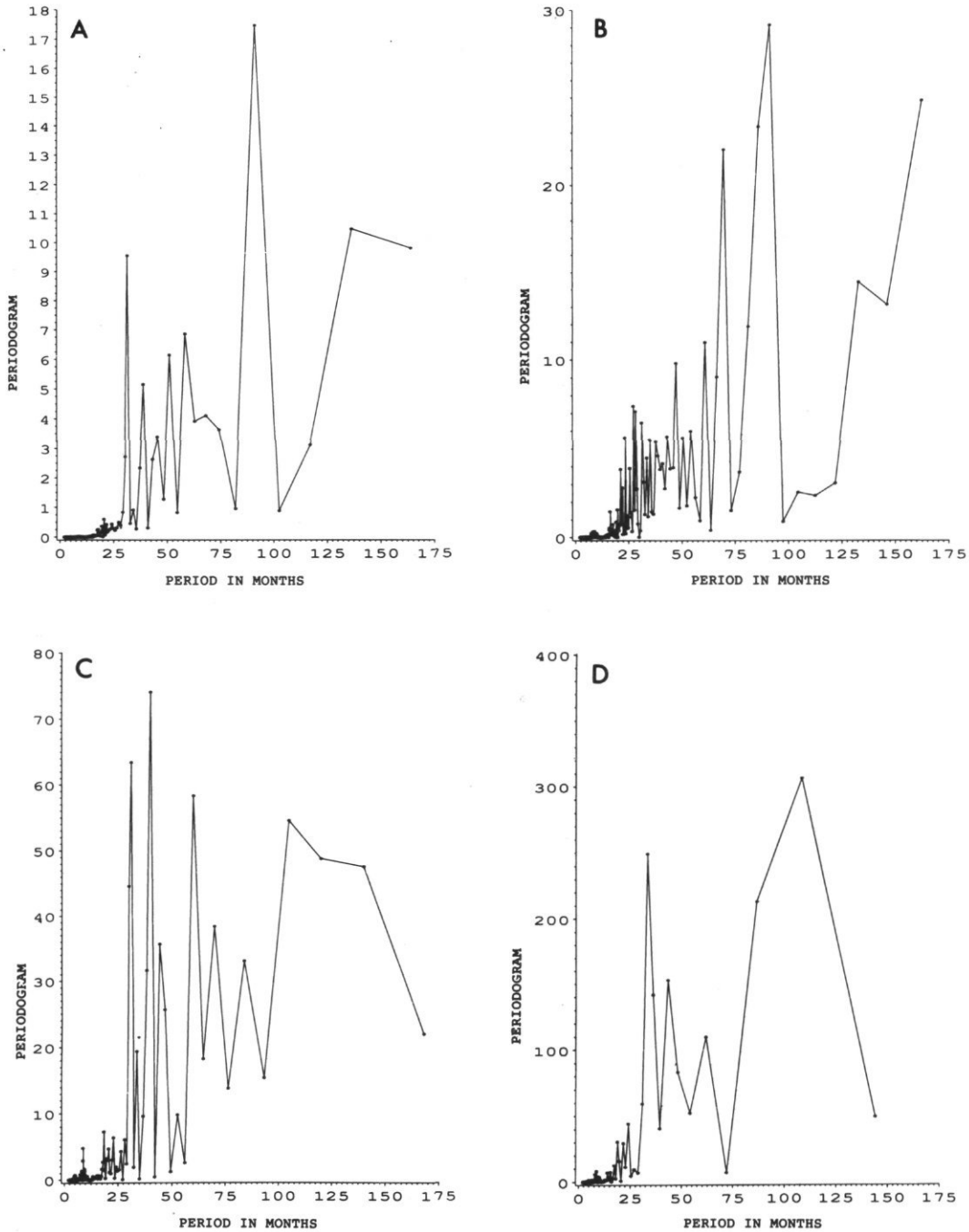

Figure 5. Periodogram for the time series from the Kola section (A, sea temperature), Ona (B, sea-surface temperature), Bjørnøya (C, air temperature), and Fruholmen (D, air pressure).

stronger warming in the north. This is similar to what was observed in the Barents Sea (Fig. 3).

Figure 5 shows four periodograms from the Fourier analysis of some time series which have first been filtered through a 12-month moving average. In all examples there are several small and large peaks. Table 2 summarizes the results from all the series we have analysed. Similar cyclic periods are found in the unfiltered series, but here the annual cycle dominates totally. The peaks listed in Table 2 are all found in at least four series, except the 17.5 years period, which shows up in two of the longest series. Because of the difficulty in precise

Table 2. Cyclic periods found by Fourier analysis. × indicates the most dominant periods in the different time series.

Section/ station	Parameter	Period (years) 2.6	3.3	5.0	7.3	8.8	11.7–13.6	17.5
Bjørnøya	AT	×	×	×		×	×	×
Bjørnøya	AP	×	×		×	×		
Fruholmen	AT	×	×	×	×			
Fruholmen	AP	×	×	×	×			
Myken	SST	×	×	×	×		×	
Ona	SST			×	×		×	
Kola	T		×		×	×	×	×
Vardø-N	T		×	×		×		
Vardø-N	S	×	×					

determination of long periodicities from relatively short series, we have put periods between 11.7 and 13.6 years together in one category. Periods of 4.2 and 6.0 years are also found in two series.

The wind-driven model has been run for the period 1970–1989. Figure 6 shows the monthly mean values of volume flux through the section from Fugløya to Bjørnøya as calculated by the model. The values vary from −1.7 Sv (1 Sv(erdrup)=10^6 m^3 s^{-1}) in March 1987 to +1.6 Sv in February 1983. It should be noted that the mean Atlantic inflow to the Barents Sea is not included in the results. The values in Figure 6 should therefore be considered as deviations from this mean, where positive values indicate an increased Atlantic inflow to the Barents Sea.

In Figure 7A, the short-term fluctuations and the seasonal signal are removed by a moving one-year average. The variations in the wind-driven flux are similar to those observed in temperature, which is clearly demonstrated by comparing with the monthly temperature anomalies in the Kola section (Fig. 7B).

Discussion

Comparison of several sections in the current system helps to reveal the advective signal in fluctuations carried by the current. The "Great Salinity Anomaly", for example, was typically advective, with a time-lag of approximately three years between the Faroe–Shetland Channel and the Spitsbergen/Barents Sea area (Dickson and Blindheim 1984; Dickson *et al.*, 1988). Also within the Barents Sea there was a delay of half a year between the western and eastern area (Loeng 1989b).

The sections in the Norwegian Sea (sections A and B in Fig. 4) go back in time only to 1978, and therefore do not cover the Great Salinity Anomaly well. The low salinity in the Svinøy section in 1978 may therefore not necessarily have been the minimum of the anomaly. It is not unlikely that it occurred earlier, since the minimum in the Faroe–Shetland Channel was observed in 1976. In the Sørkapp section (Fig. 4C) the minimum was observed in 1979 (Dickson and Blindheim, 1984) and, with regard to salinity, Figure 4B indicates the same year for

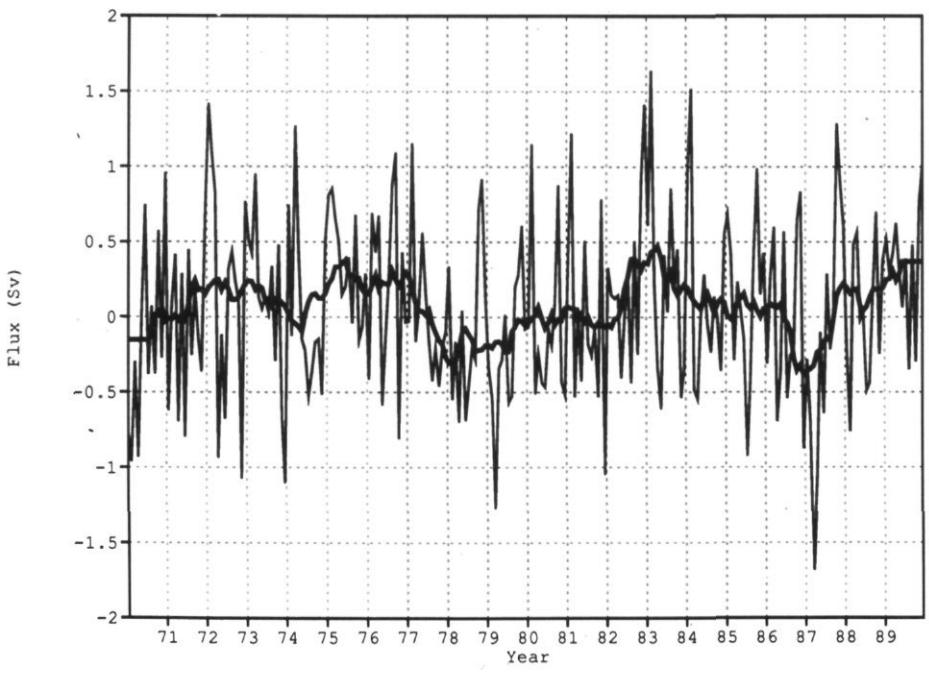

Figure 6. Modelled wind-driven volume flux through the Fugløya–Bjørnøya section during the period 1970–1989. The thin line represents monthly means, while the heavy line is a one-year moving average.

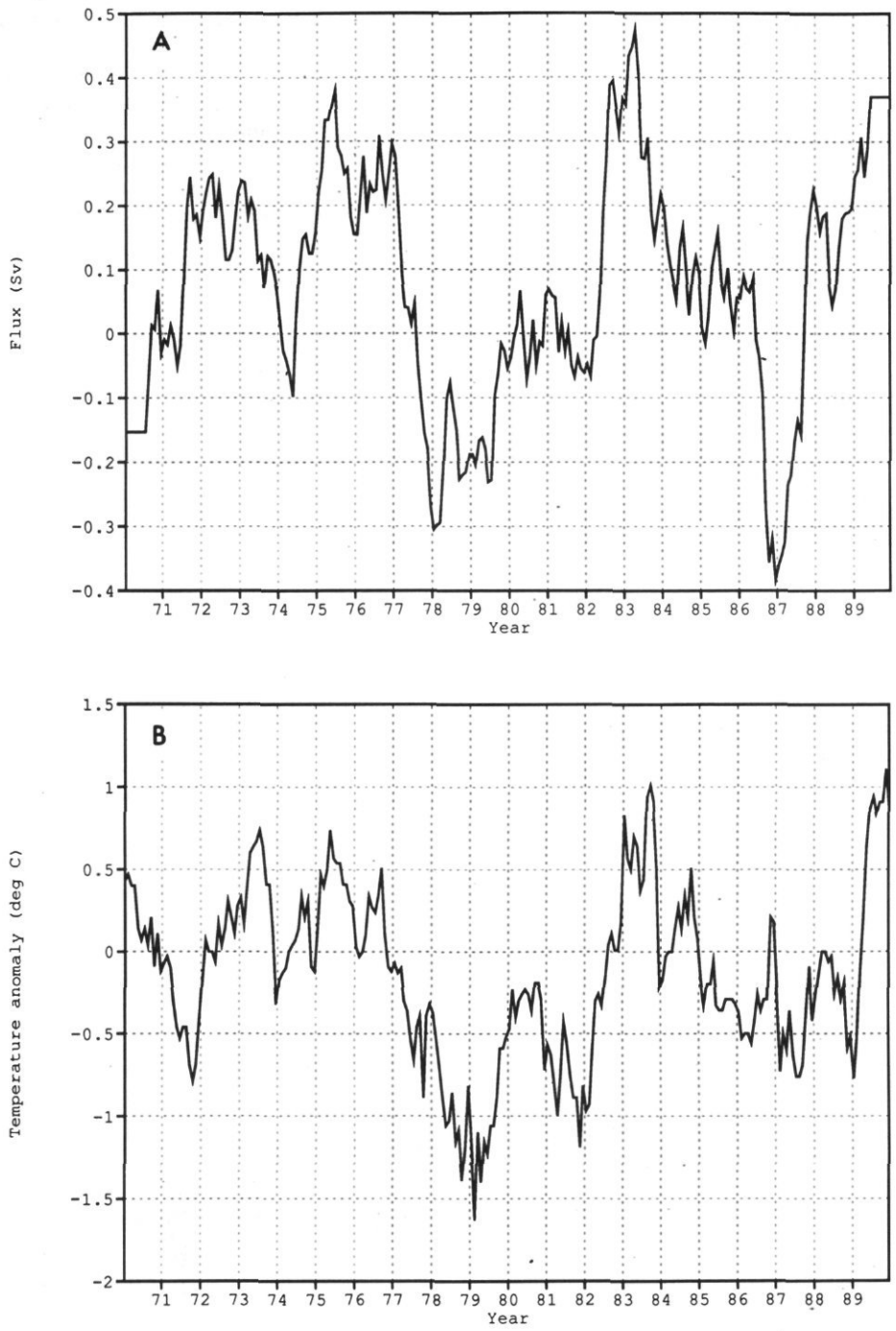

Figure 7. (A) One-year moving average of modelled flux through the Bjørnøya–Fugløya section. (The vertical axis is rescaled from Figure 6.) (B) Measured monthly temperature anomalies in the Kola section for the same period.

the minimum off Lofoten. The reason for this is probably the low observation frequency in these sections. A time-lag would probably appear clearly if observations were more frequent, as in the Barents Sea (Fig. 3). In the Bjørnøya section the salinity minimum was observed in late 1978, while it was observed about six months later in the Vardø-N section.

The variations during the 1980s were more complex. Although there were similarities in the sections, the trends also showed differences. The 1983 temperature

maximum in the Svinøy section was also observed further north and there were indications of a time-lag. In salinity this peak was not clearly reflected northwards. It was repeated in the Gimsøy section, but to less distinction in relation to the Svinøy section. At Spitsbergen it was hardly noticeable. As observed in the Sørkapp section, salinity in this area peaked in 1980/1981 when the associated temperatures also were high. Both the temperature and the salinity signal of this event showed a southward decreasing magnitude; this was not observed in the Barents Sea sections (Fig. 3). The model computations (Fig. 7A) indicate a flux of about average into the Barents Sea in this period. It therefore seems reasonable to conclude that most of the Atlantic water went into the West Spitsbergen Current during 1980 and 1981, possibly carried by the western branch of the Norwegian Atlantic Current in the Norwegian Sea. The westward distribution of 0-group fish in these years is in support of this conclusion (Anon., 1980, 1981).

The maximum in 1983, on the other hand, is clearer in the Barents Sea than at Spitsbergen. During this period the model also indicates a sharply increased flux into the Barents Sea. Although there is a fairly clear temperature maximum at Spitsbergen, too, it cannot be identified in salinity. The reason for this is not evident. A possible explanation may be that the waters from the western branches of the Norwegian Atlantic Current were dominant in the West Spitsbergen Current during this period. In general, however, there is fairly good agreement with the model results. When the atmospheric forces drive much water into the Barents Sea, this is reflected by a weak Atlantic signal in the West Spitsbergen Current. In opposite cases, as for the 1980/1981 peak in the Sørkapp section, the Atlantic signal is stronger in the West Spitsbergen Current and weaker in the Barents Sea.

The volume transport of the current system is not taken into consideration. If this was better known, mismatches could possibly be better explained. This would also help in explaining the temperature/salinity increase during 1989/1990, which started more or less simultaneously in the southern Norwegian Sea, the West Spitsbergen Current and the Barents Sea. The weak advective and strong temperature signal in this event may suggest that the locally reduced heat loss was of greater importance than an intensification of the inflow.

Another process of importance may be the formation and outflow of dense bottom water from the eastern basin in the Barents Sea. During the winter, bottom water is formed as a result of cooling and ice formation, a process described in more detail by Midttun (1985). The bottom water formed over the Central Bank and the shelf area of Novaya Zemlya flows into the eastern basin, which occasionally may be almost filled with this water. Most of this bottom water leaves the Barents Sea through the strait between Novaya Zemlya and Frans Josef Land and the outflow has to be replaced by inflowing water from the west. In volume, the outflow may vary considerably from year to year; likewise the corresponding inflow. Major outflows occurred in 1982/1983 and in 1989/1990, when there was a rapid temperature increase in the Barents Sea (Loeng, 1991).

The overall patterns in the model flux and the Kola series (Fig. 7) are very similar. There are at least three factors influencing the temperature in the Kola sections. These are: the volume flux of the inflowing Atlantic water, the temperature of this water mass, and the local heat flux to the atmosphere. The model considers only a part of the first factor, the part that is caused by local wind forcing.

The agreement between the series does not necessarily imply that it is the local wind conditions that have the greatest influence on the temperature. Another interpretation is that the agreement indicates the dependency between the above three factors. Warm periods are characterized by low air pressure over the Barents Sea giving a cyclonic circulation with southwesterly winds over the area. This causes inflow to the Barents Sea to increase. In cold periods the air pressure is higher giving northeasterly cold winds with both decreased inflow and enhanced cooling of the Atlantic water. These kinds of feedback mechanisms in the Barents Sea climate have been discussed in more detail by Ådlandsvik and Loeng (1991). Ikeda (1990) and Mysak *et al.* (1990) have discussed the importance of feedback mechanisms in a larger context.

There are, however, discrepancies between the series in Figure 7. The most noteworthy is the flux minimum in winter 1987, which is not reflected in a similar way in the Kola data; in fact the Kola series has a local maximum in the same period. This maximum in the temperature anomaly may be due to a late summer in 1986 with temperature maximum in November. The non-smoothed model results (Fig. 6) show a very marked minimum in the volume flux in March 1987. It is likely that the inflow of Atlantic water was only temporarily blocked, and returned to normal or high flux when the winds calmed. Such an increase in the inflow would not be captured in this purely wind-driven model.

The Fourier analysis revealed a large complexity in the periodicity of all the time series. Since most of the peaks occurred in different time series of different length, we consider them as real periods. Periods of three to five years without any closer analysis have been suggested earlier (Kissler, 1934; Lunde, 1965). Our periods of 2.6, 3.3, and 5.0 years concur with these suggestions. The medium periods of 7.3 and 8.8 years have not been mentioned earlier for this area. Because of the lengths of the series it is difficult to define the exact lengths of the longest periods. It is therefore difficult to state whether there is more than one period between 11.7 and 13.6 years. The minimum of this period is very close to the solar activity cycle of 11 years proposed for ice and sea temperature variability by Bochkov (1976). Further, Mysak *et al.* (1990) indicate a climatic cycle for

the Greenland Sea of 10–15 years. It should also be mentioned that strong year classes of cod in the Barents Sea have occurred every 13 years during the period since 1937 (Loeng 1989a). The longest period detected was 17.5 years. However, this is double 8.8 years, and may therefore be a harmonic component.

With principal cycles from 7.3 to 13.6 years, the Ona series differs from the others. The Ona series is by far the longest, but this is not the explanation for the difference. When analysing the series from 1920 to 1990, we got the same results as for the whole series. It therefore seems that there is a real difference in the principal periodicities between Ona and the other stations.

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Sources of fresh water in the Iceland Sea and the mechanisms governing its interannual variability

Steingrímur Jónsson

Jónsson, S. 1992. Sources of fresh water in the Iceland Sea and the mechanisms governing its interannual variability. – ICES mar. Sci. Symp., 195: 62–67.

Hydrographic data from a section extending from Langanes to Jan Mayen are used to study the seasonal and interannual variability in the water-mass structure of the Iceland Sea. The freshwater thickness in the upper 150 m is found to be a good indicator of the conditions in the Iceland Sea. The freshwater runoff from Iceland does not contribute much to the freshwater content in the Iceland Sea. The evaporation–precipitation cycle is shown to be an unlikely source of the observed interannual variability of the freshwater thickness. The only remaining source of the interannual variability is the East Greenland Current, which carries with it large amounts of fresh water from the Arctic Ocean. It is suggested that the mechanism by which fresh water is released from the East Greenland Current and allowed to flow into the Iceland Sea is governed by the wind stress curl over the area, and some evidence for this is shown. The possible consequences of this mechanism for the water mass transformations in the Greenland Sea are discussed, in particular the reduced deep water formation observed there during the 1980s.

Steingrímur Jónsson: Marine Research Institute, PO Box 1390, Skúlagata 4, 121 Reykjavík, Iceland.

Introduction

The Iceland Sea (Fig. 1) covers the area bounded by Iceland, Greenland, and Jan Mayen. It is about 500 000 km^2 in area and 400 000 km^3 in volume (Aagaard *et al.*, 1985). Ignoring nearshore regions, the area is 350 000 km^2 and the volume is 395 000 km^3 giving an average depth of 1130 m (Swift, 1980).

The circulation in the Iceland Sea, according to Stefánsson (1962), is shown in Figure 1. In the eastern basin it consists of a cyclonic gyre centred at 68.5°N 12°W. This gyre is bordered by the East Icelandic Current in the south, and to the west by the East Greenland Current. To the east, Atlantic water is carried northwards and then flows into the Iceland Sea south of Jan Mayen. South of the East Icelandic Current the North Icelandic Irminger Current flows along the Icelandic shelf carrying with it water of Atlantic origin. This current system was also revealed by the dynamic topography calculated by Swift (1980), using hydrographic data covering the whole Iceland Sea.

It was shown by Aagaard (1970), and later by Jónsson (1991), that within the framework of Sverdrup dynamics the wind stress curl over the Nordic Seas is capable of driving considerable transports.

Legutke (1991) used a three-dimensional numerical model to study the vorticity balance in the Nordic Seas. The main terms in the vorticity balance were the bottom pressure torque and viscous terms, and only in regions of negligible topographic gradients, e.g. in the flat basins, was an approximate flat bottom Sverdrup balance found. It thus seems that the important factors determining the flow in the Nordic Seas are the bottom topography and the forcing, which is in agreement with the topographic Sverdrup balance.

In the Nordic Seas deep convection is restricted to the cyclonic gyres of the Iceland and Greenland seas because of their very low stratification. The temperature gradient built up during summer in the upper ocean is easily broken down by the winter cooling. Apart from this temperature gradient the stratification depends on a slightly reduced salinity of the surface layer. Aagaard and Carmack (1989) have estimated the freshwater content over an area of 140 000 km^2 of the Iceland Sea gyre using a reference salinity of 34.93 and found it to correspond to a freshwater thickness of 0.61 m. About half of this they account for by an excess of precipitation over evaporation, but the rest is accounted for by advection of low salinity water from the East Greenland Current. Aagaard and Carmack (1989) conclude that

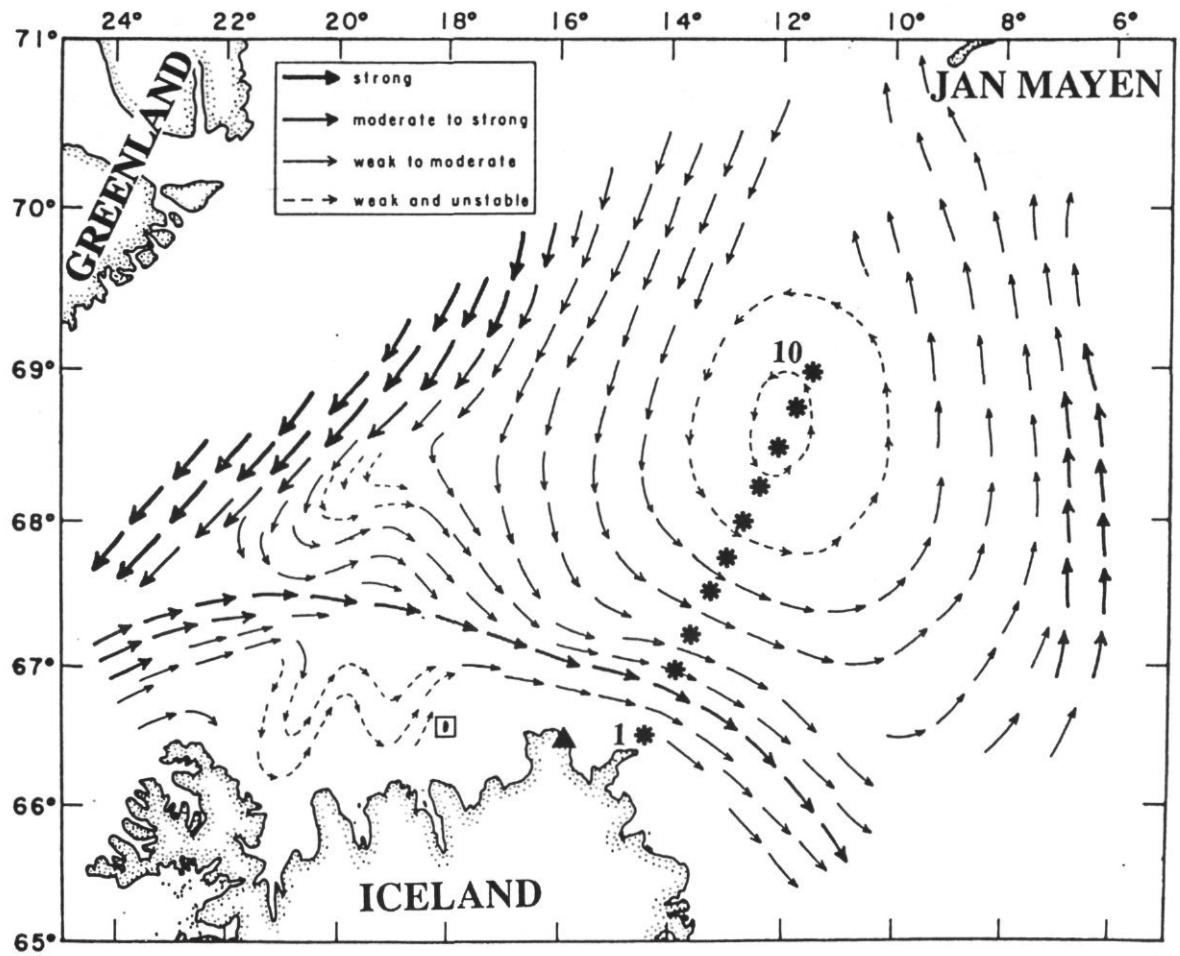

Figure 1. The circulation in the Iceland Sea (from Stefánsson, 1962). Also shown are the stations on the Langanes–Jan Mayen section. The positions of the meteorological stations Grímsey (square) and Raufarhöfn (triangle) are also indicated.

the upper layers of the East Greenland Current are nearly isolated from the interior convective gyres, and if the freshwater flux from the boundary current were to increase slightly, convection would probably cease. Thus it is the release of fresh water from the boundary current that controls the ventilation of the Greenland and Iceland seas, and it is therefore of great interest to try to understand the mechanism behind this control. Foldvik *et al.* (1988) have shown that the turbulent heat flux across the Polar front in Fram Strait is small. However, their shallowest instrument was at 78 m depth, so they did not cover the wind-driven Ekman layer.

It is the purpose of this paper to investigate the interannual hydrographic variability in the eastern Iceland Sea, and its relation to the variability of the wind stress curl over the area. In particular, we will investigate the role of the wind stress curl in controlling the flux of low salinity water from the boundary current into the convective gyres, thus governing the stratification there.

Material

Hydrographic data from a section from Langanes towards Jan Mayen (Fig. 1) obtained by the Marine Research Institute in Reykjavík will be used. This section has been occupied frequently since 1950, and more regularly, four times each year, since 1970. Stations 1 and 2 are positioned on the shelf, while station 3 lies over the continental slope; the rest of the stations lie above the flat Iceland Plateau, which is about 1900 m deep.

Wind data from the hindcast database of the Norwegian Meteorological Institute are used. The database and the method used for calculating the wind stress curl have been described by Jónsson (1991).

In order to investigate the role of the hydrological cycle of precipitation and evaporation on the freshwater content of the Iceland Sea, data on precipitation from Grímsey and Raufarhöfn (Fig. 1) (Anon., 1971–1989) have been used. The Norwegian Meteorological Institute provided precipitation data from Jan Mayen.

Results

The water masses present in a section from Langanes to Jan Mayen can be seen in Figure 2, which is from February 1971. The hydrographic conditions in the East Icelandic Current, and their variability, have been studied by Malmberg (1972, 1984). The North Icelandic Winter Water (NIWW) is present closest to the shore. This water mass was defined by Stefánsson (1962) as having a temperature of 2–3°C and a salinity of 34.85–34.90, but its composition can vary a great deal. It is formed during winter by the mixing of Atlantic and Arctic waters. From the surface down to about 150 m the central part of the Iceland Sea is covered by a homogeneous water mass, the upper Arctic Intermediate Water (uAIW). This water is formed by convection during winter cooling and has temperatures less than zero, but the salinity varies from one year to the next, but usually it is less than 34.8 (Malmberg, 1984). At about 150 m depth there is usually a relatively sharp halocline. Below the uAIW there is a temperature maximum and sometimes a salinity maximum. This water mass, the 1AIW, was defined by Swift and Aagaard (1981) to have temperatures above 0°C and salinity above 34.9. It can always be identified by a temperature maximum, but this is frequently lower than 0°C. There is often no salinity maximum, and the salinity at the temperature maximum is often below 34.9. The 1AIW is an advective feature, and has its origin in the Atlantic water that sinks below the East Greenland Current off Spitsbergen and also in the Atlantic water that enters the Iceland Sea south of Jan Mayen (Swift, 1986). The latter water mass can be seen in the figure as a salinity and temperature maximum off Jan Mayen. Below the 1AIW, Norwegian Sea Deep Water (NSDW) and Arctic Ocean Deep Water (AODW) fill the basin. The main emphasis here will be on the uAIW, which shows the greatest seasonal and interannual variability and is in direct contact with the atmosphere.

As an example of the seasonal variation, data from 1987 at various times at station 6 are shown in Tables 1 and 2.

In February the water column is homogeneous in both temperature and salinity down to roughly 100 m. Below that the temperature rises, reaching a maximum at 300 m before decreasing again. Below 100 m the salinity gradually increases, but there is no extreme. In June the summer warming of the surface layers has started, and a

Table 1. T at station 6 at different times during 1987.

Depth (m)	Jan 25	Feb 16	Jun 6	Sep 7	Oct 18
0	−0.36	−0.70	3.65	7.64	2.85
10	−0.35	—	2.48	7.59	—
20	−0.32	−0.69	1.93	7.59	2.84
30	−0.32	—	1.60	6.93	—
50	−0.30	−0.68	0.56	0.07	2.84
75	−0.29	−0.68	0.29	−0.42	−0.32
100	−0.29	−0.60	−0.60	−0.65	−0.26
150	−0.20	−0.20	−0.71	−0.40	−0.43
200	−0.15	−0.20	−0.14	−0.20	−0.17
300	−0.19	−0.12	−0.07	−0.16	−0.15
400	−0.26	−0.23	−0.25	−0.29	−0.24
500	−0.30	—	−0.33	−0.41	−0.33

Table 2. S at station 6 at different times during 1987.

Depth (m)	Jan 25	Feb 16	Jun 6	Sep 7	Oct 18
0	34.66	34.62	34.64	34.48	34.66
10	34.64	—	34.64	34.50	—
20	34.65	34.63	34.65	34.50	34.66
30	34.65	—	34.66	34.53	—
50	34.65	34.63	34.69	34.71	34.66
75	34.66	34.64	34.71	34.75	34.74
100	34.66	34.65	34.71	34.78	34.80
150	34.87	34.78	34.78	34.84	34.84
200	34.89	34.86	34.88	34.88	34.88
300	34.89	34.90	34.90	34.91	34.90
400	34.90	34.90	34.90	34.90	34.91
500	34.91	34.91	34.90	34.91	34.91

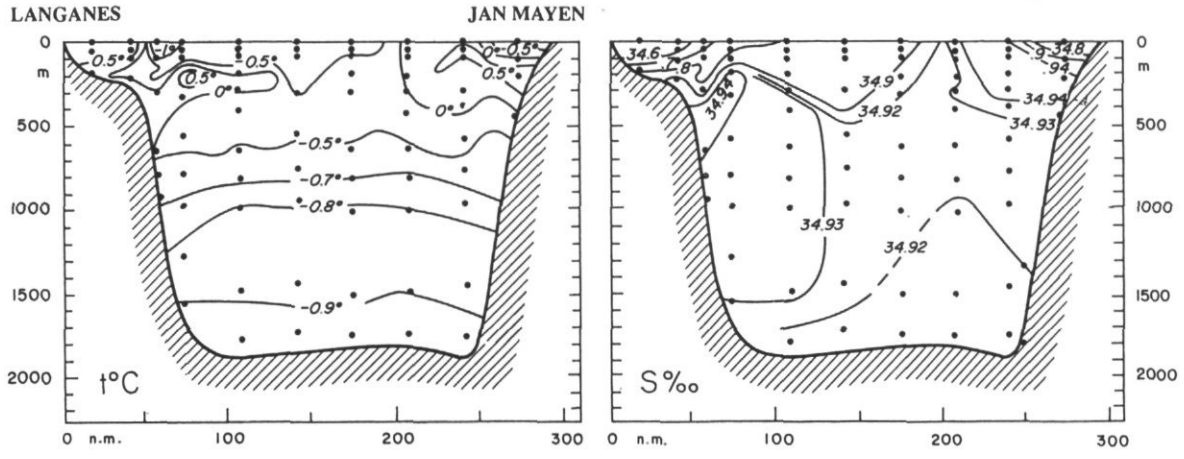

Figure 2. Temperature and salinity in the Langanes–Jan Mayen section in February 1971 (from Malmberg, 1983).

Table 3. Correlation between the freshwater thickness at different stations.

	Station no.							
Station no.	1	2	3	4	5	6	8	10
1	1.00	0.83	0.76	0.54	0.38	0.51	0.10	0.67
2		1.00	0.74	0.44	0.24	0.37	−0.21	0.55
3			1.00	0.73	0.70	0.79	0.40	0.70
4				1.00	0.83	0.92	0.53	0.78
5					1.00	0.88	0.61	0.72
6						1.00	0.93	0.97
8							1.00	0.43
10								1.00

thermocline develops. This warming continues until September and the surface layers reach temperatures of 6–8°C. The surface waters also acquire a weak salinity stratification during the summer. Because the seasonal variations are very weak below 150 m depth the freshwater thickness of the upper 150 m of the water column relative to a constant salinity of 34.90 should give a good indication of the conditions in the Iceland Sea; this has been calculated for each station.

In Table 3 the correlation between the freshwater content in spring at the stations on the Langanes–Jan Mayen section is shown. Because of sparse data, stations 7 and 9 are not included. This shows that the variability north of station 3 is quite coherent, while the variability at the shallowest stations is decoupled from that at the deep stations. This is probably due to the fact that the salinity at the shallowest stations is influenced by runoff from land, and the supply of high salinity water from the Irminger Current that flows into the Iceland Sea at the northwest corner of Iceland. This correlation structure is in accordance with the results of Swift and Aagaard (1981) that the Atlantic Water that enters the southwestern Iceland Sea is closely confined to the north Icelandic shelf and then flows around Iceland and out of the Iceland Sea to the southeast, without entering or otherwise contributing to the central Iceland Sea.

In an attempt to estimate the influence of the evaporation–precipitation cycle on the stratification in the Iceland Sea, data on precipitation from Grímsey have been examined. Grímsey is a small island about 40 km off the north coast, and conditions there should therefore be representative of conditions in the open ocean. The mean annual precipitation for the years 1971–1989 is 655 mm. In Raufarhöfn and Jan Mayen the average precipitations for the same period are 735 and 734 mm respectively. The mean annual value for the evaporation over the Iceland Sea is about 500 mm (Hopkins, 1988), so it is only slightly less than the precipitation. One has to bear in mind, however, the fact that measurements of evaporation are rare and inaccurate and no long time series exists. We can conclude that the excess of precipitation over evaporation is probably less than 300 mm. If one assumes that the residence time of the surface and intermediate waters in the Iceland Sea is about one year, which is probably too large (Swift and Aagaard, 1981), the precipitation–evaporation cycle would probably contribute less than 0.3 m to the freshwater thickness there. Any fresh water above this amount would have to originate from fresher water being advected from the bordering currents, the only possibility being the East Greenland Current. Interannual variations in the precipitation in Grímsey are shown in Figure 3. The variations are rather small, with the standard deviation being 139 mm. Furthermore, there is a negative correlation between the freshwater content in spring at station 4 and the precipitation averaged over a one-year period prior to the salinity measurement. Thus most of the interannual variability of the freshwater content must originate from advection from outside, i.e. from the East Greenland Current.

The mean of the wind stress curl over the Nordic Seas for the period 1955–1987 is shown in Figure 4. It is characterized by positive values over most of the area, and only close to the Greenland coast and around the Faroe Islands are negative values observed. The corresponding flat-bottom Sverdrup balance is shown in Figure 5. We note the large transports possible within this framework, the maximum being 35 Sv. Most of this transport occurs within the Nordic Seas and relatively small amounts are exchanged with the adjacent seas. It is therefore not unreasonable to assume that the wind stress curl has a major effect on the circulation in the gyres.

The mean wind stress curl that is positive gives rise to an upward-directed Ekman pumping and a cylonic circulation in the gyres as is observed. This Ekman pumping causes a divergence at the surface, thus removing the freshest and lightest water and replacing it with heavier and saltier water at a somewhat deeper level. This mechanism, which pumps salty water into the area and fresh surface water out of it, would effectively reduce the stability of the water column in the gyres, making deep convection more likely to take place. If the wind stress curl were to be weakened or reversed, this would lead to an increased freshwater inflow from the boundaries and to an increased freshwater content in the Iceland Sea. This would eventually prevent deep convection. In order to investigate this, the freshwater thickness at

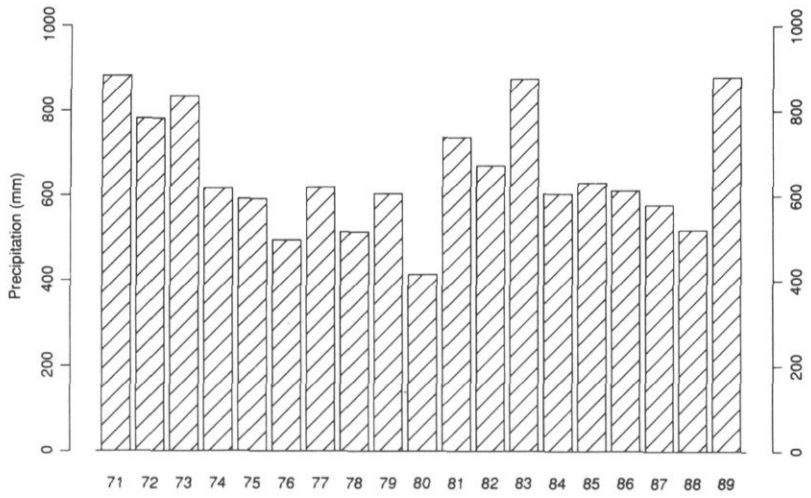

Figure 3. Interannual variations in precipitation in Grímsey 1971–1989.

station 4 in spring as a function of the wind stress curl averaged in space over the Iceland Sea gyre and in time over a one-year period prior to the salinity measurement is shown in Figure 6. Jónsson (1991) showed that the wind stress curl is very weak during summer, while it is very strong from October to April. The wind stress curl average is therefore representative of the winter prior to the salinity measurement, and one can look at the freshwater thickness in spring as the result of the working of the wind on the circulation during the winter. There is an obvious trend for large freshwater content to be associated with small wind stress curl. The extreme ice years in North Icelandic waters, 1965, 1968 and 1969, are all characterized by very small wind stress curl, while for the warm, high salinity period prior to 1965, the wind stress curl is very high.

Discussion

It has been argued here that the freshwater content of the Iceland Sea depends on the conditions in the atmosphere. Since the amount of fresh water is crucial for the water mass transformations occurring there, it is of interest to determine whether such influence is also to be

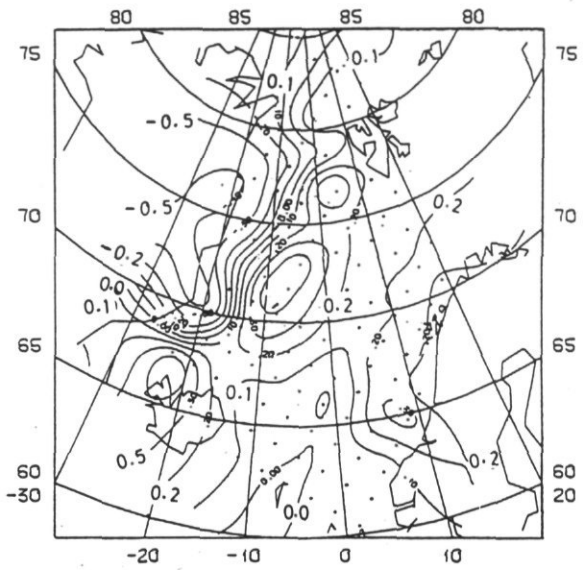

Figure 4. Mean wind stress curl over the Nordic Seas for the period 1955–1987. The unit is Pa/1000 km and the contour interval is 0.1 Pa/1000 km (from Jónsson, 1991).

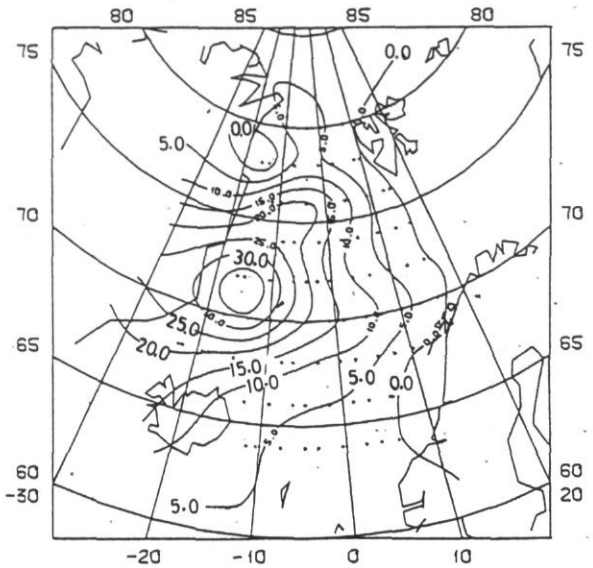

Figure 5. The integrated Sverdrup transport corresponding to the mean wind stress curl in Figure 4. The unit is Sv and the contour interval is 5 Sv (from Jónsson, 1991).

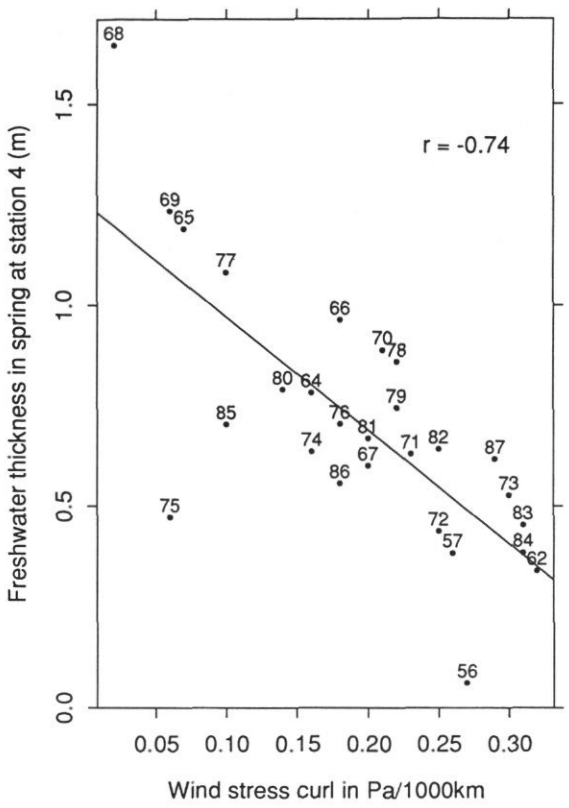

Figure 6. The freshwater thickness at station 4 as a function of the wind stress curl over the Iceland Sea gyre. The number at each point is the year-1900 of the salinity measurement.

expected elsewhere. This would be especially relevant in the case of the Greenland Sea, where bottom water formation occurs. It has been reported that there has been no bottom water production there since 1982 at least (GSP Group, 1990). Unfortunately, there exists no standard section in the Greenland Sea that is occupied regularly, as in the Iceland Sea, and the database is much more widespread. In this connection it is therefore interesting to see whether there has been a change in the atmospheric circulation during the 1980s compared with previous years. This has been investigated by Jónsson (1991), who showed that an empirical orthogonal function (EOF) analysis of the wind stress curl field revealed that in the 1980s it showed anomalous behaviour. The amplitude of the first mode that describes the behaviour of the large cell over the central Greenland Sea (Fig. 4) is weaker after 1980 than in previous years. Perhaps it is this reduction in wind stress curl that has led to a freshening of the Greenland Sea and thus limited the convection there in the 1980s.

The arguments put forward here point to the wind stress curl as being a major factor controlling the hydrographic variability in the Arctic Domain of the Nordic Seas. It is therefore important that the role of the wind field over the Nordic Seas is studied further, especially its role in controlling the water mass transformation there.

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Decadal variations of ocean climate in the Norwegian Sea observed at Ocean Station "Mike" (66°N 2°E)

Tor Gammelsrød, Svein Østerhus, and Øystein Godøy

Gammelsrød, T., Østerhus, S., and Godøy, Ø. 1992. Decadal variations of ocean climate in the Norwegian Sea observed at Ocean Station "Mike" (66°N 2°E). – ICES mar. Sci. Symp., 195: 68–75.

Temperature and salinity observations obtained at "Mike" since 1948 are presented. Average vertical profiles are established. The evolution of the vertical structure of the "Great Salinity Anomaly", when it passed the Norwegian Sea in the late 1970s, is discussed. A recent marked temperature increase at 2000 m depth is related to the cessation of the formation of deep water in the Greenland Sea during the 1980s.

Tor Gammelsrød, Svein Østerhus, and Øystein Godøy: Geofysisk Institutt, Universitetet i Bergen, 5007 Norway.

Introduction

The weather ship station "Mike" in the Norwegian Sea (66°N 2°E) has been occupied since 1948. Besides the standard meteorological routines, an oceanographic observation program has been carried out. This program consists of daily Nansen casts, including one deep station every week. The water depth is more than 2000 m in this position, so the "Mike" data constitute the longest and most complete time series from the deep ocean.

The "Mike" time series is of particular value for the study of climate variations on decadal time scales. According to the IPCC report (Houghton *et al.*, 1990), two of the basic reasons that climate fluctuations are only poorly understood are the lack of both homogeneous time series and oceanic observations. Here we meet these two needs simultaneously. The position in the Norwegian Sea (Fig. 1) is also strategic from the point of view of monitoring climate fluctuations: the upper few hundred meters are dominated by the warm, saline water of the Norwegian Atlantic current, which is the most important contribution of oceanic heat transport towards high latitudes in the Northern hemisphere, and at great depths the water is influenced by the bottom water formed in the adjacent Greenland Sea. So variations in key climate processes like poleward heat transport and bottom water formation may be recorded at "Mike".

Data and methods

Water samples, taken with Nansen bottles equipped with reversing thermometers, are analysed for oxygen content onboard and salinity concentration ashore after each cruise, which usually lasts one month. The observation depths are corrected using unprotected thermometers. Temperature and salinity data are presented here.

Mean values near the standard observation depths were constructed by averaging all the observations within a certain depth range. Because of the small vertical gradient at greater depths, this depth range was chosen from about 5 m in the upper layer to about 50 m near the bottom. The results are given in Table 1.

There are more observations from the upper layers than from great depths. However, as the seasonal variability is confined to the upper 300 m, and the variations at great depths are slow, the observation frequencies are adequate for studies throughout the water column.

Monthly averages were computed at the standard depths. Mean monthly averages for the whole observation period were calculated to study seasonal variations. These were subtracted from the monthly averages to form time series of anomalies. If observations from one month were missing we used the mean of the previous and subsequent months. In cases when several consecutive monthly values were missing, the

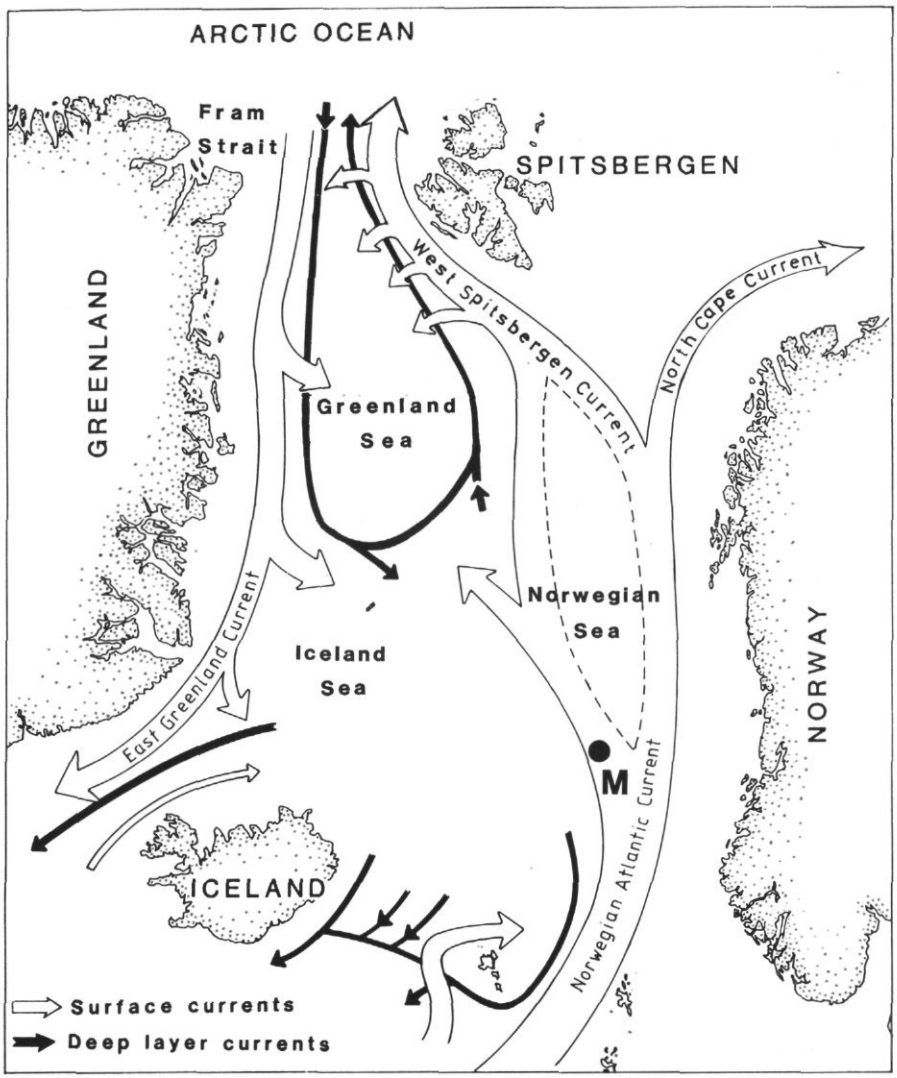


Figure 1. Main current systems (schematic) in the Nordic Seas. The weathership station position "M" is also shown.

Table 1.

Depth (m)	s.d. (m)	Temperature (°C)	s.d. (°C)	Number of observations	Salinity (psu)	s.d. (psu)	Number of observations
0	1	8.306	1.9823	8017	35.180	0.0751	6939
10	0	8.213	1.9282	3385	35.146	0.0670	2936
25	0	7.997	1.6104	3069	35.150	0.0658	2851
49	2	7.443	1.1131	7410	35.172	0.0670	6883
74	2	7.086	0.8424	3241	35.183	0.0696	3211
99	2	6.887	0.7722	3079	35.180	0.0668	3075
149	4	6.487	0.8625	5747	35.163	0.0724	5539
198	3	6.173	0.8946	1657	35.144	0.0815	1758
298	7	4.696	1.4237	4071	35.057	0.0846	3878
397	6	2.988	1.3508	2984	34.990	0.0631	2964
496	8	1.264	0.7670	1243	34.952	0.0341	1307
595	6	0.496	0.5419	4319	34.932	0.0224	4219
796	11	−0.305	0.1423	1048	34.922	0.0146	1097
991	12	−0.601	0.0488	3514	34.919	0.0060	3841
1191	15	−0.742	0.0541	1184	34.919	0.0061	905
1493	14	−0.869	0.0360	983	34.919	0.0062	841
1996	36	−0.945	0.0208	633	34.918	0.0061	537

monthly averages for the actual months the year before and after were used.

To study long-term trends we filtered the monthly anomalies using a squared Butterworth filter with a 1/4 power point at about four years (Roberts and Roberts, 1978).

Results

Mean vertical profiles

Average vertical profiles with the associated standard deviations are shown in Figure 2 (Table 1). The seasonal variations are confined to the upper 300 m (Gammelsrød and Holm, 1984) and explain the high variability in the surface layer. However, even from 300 m down to about 800 m the variability is high. This is probably due to meanderings and changes in the position of the Norwegian Atlantic Current (Fig.1).

The stratification in the Norwegian Atlantic Current is dominated by the temperature. The salinity profile tends to destabilize the water column below the salinity maximum at 75 m depth. The salinity minimum at the surface layer is due to precipitation and the influence of fresh water from the Norwegian Coastal Current (Helland, 1963).

A salinity minimum, due to the influence of Arctic Intermediate Water, is usually found in the Norwegian Sea between 500 m and 800 m depth (Blindheim, 1990). This minimum does not show up in the average profiles (Fig. 2) because the vertical resolution of the Nansen casts is too poor (Table 1), and it is also smoothed out in the averaging process. However, we were able to see the salinity minimum in many of the individual profiles, and it also appears in the average minimum profile (computed by subtracting the standard deviation from the mean), between 600 m and 800 m depth (Fig. 2).

Time series

Monthly temperature and salinity anomalies at the surface layer, represented by the 50 m depth, and at 1500 m and 2000 m are shown in Figure 3. Note that the ordinates are different for the surface and the two deep time series.

The 50 m is representative of the upper 300 m, as the variations here are more or less parallel (Gammelsrød and Holm, 1984). We notice the warm saline period in the early 1960s, as well as the "Great Salinity Anomaly" (Dickson *et al.*, 1988) in the second half of the 1970s. This feature is believed to be caused by an increased freshwater transport in the East Greenland Current. Mysak and Power (1991) point out that the "Great Salinity Anomaly" was preceded by an above average runoff from North America into the western Arctic. This feature was traced in a cyclonic path in the northern North Atlantic from the Iceland Sea in 1968, passing the Norwegian Sea about 10 years later (Dickson *et al.*, 1988).

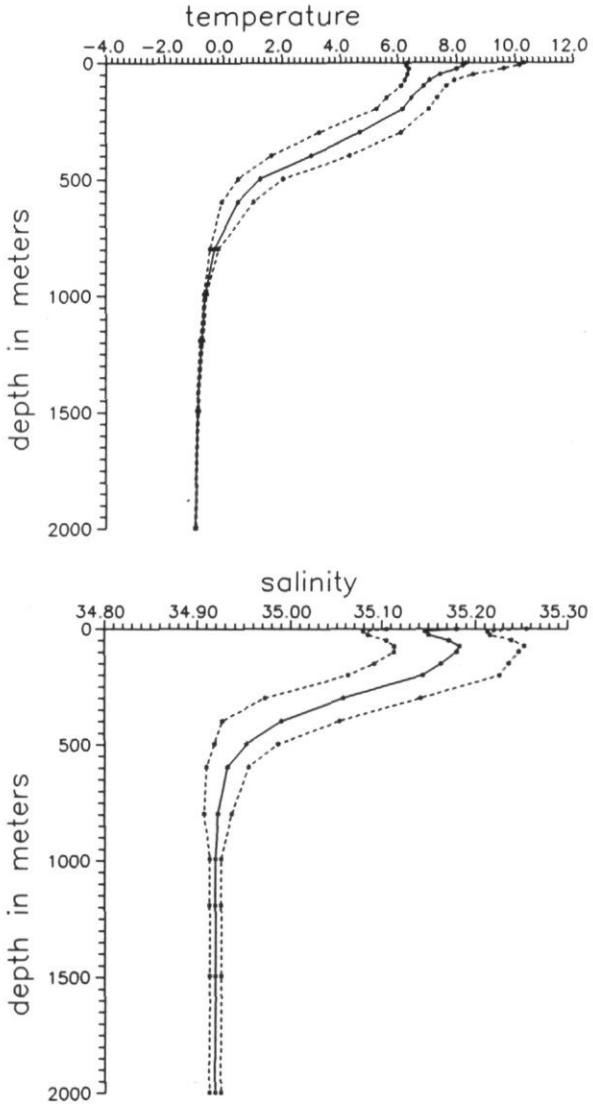


Figure 2. Average vertical profiles of temperature and salinity with associated standard deviations. The numbers are given in Table 1.

The vertical structure and the time evolution of the "Great Salinity Anomaly" in the Norwegian Sea are seen in the isopleth diagrams (Fig. 4b). The relative fresh water of the "Great Salinity Anomaly" is accompanied by a low water temperature (Fig. 4a), but not enough to compensate for a deficit of the water density in the order of 0.05 σ units (Fig. 4c). It may be observed (Fig. 4b) that the "Great Salinity Anomaly" arrived at the upper 200 m at "Mike" in 1975, was gradually deepening down to 500 m in 1979, but disappeared in the whole water column simultaneously in 1980.

In the deep ocean (1500 m and 2000 m, Fig. 3) the amplitudes of the fluctuations are small. Even at these depths the temperature during the 1960s was above average, while the salinity was slightly below average.

Recent trends

In the surface layer the temperature has leveled out at about average since the "Great Salinity Anomaly". The salinity seems to have a small negative trend (Fig. 3).

In the deep ocean the temperature remained constant below the average at 1500 m during the 1980s. In contrast there was a definitive warming trend during the last decade at the 2000 m level. The temperature at 2000 m reached its absolute minimum value in 1981, increasing to the maximum of the whole period in 1990; the temperature rise was 0.07 ± 0.02°C.

The deep ocean salinities seem to have been above average during the last decade.

Discussion

Most of the bottom water in the Nordic Seas is believed to be formed by deep convection in the Greenland Sea (Helland-Hansen and Nansen, 1909; Mosby, 1970; Clarke *et al.*, 1990). The Greenland Sea Deep Water (GSDW) is observed to be colder, but also slightly less saline than adjacent bottom water masses (Aagaard *et al.*, 1985).

Decadal variations of the abyss temperature in the Greenland Sea (Clarke *et al.*, 1990) seem to be consistent with the variations at 2000 m at "Mike" (Fig. 3a). Using chlorofluorocarbon (CFC) as a tracer and a box model, Schlosser *et al.* (1991) concluded that the production of GSDW was reduced from about 0.5 Sv to about 0.1 Sv some time between 1978 and 1982.

Swift and Koltermann (1988) argue that Norwegian Sea Deep Water (NSDW) is formed by a mixing of about equal parts of GSDW and Eurasian Basin Deep Water (EBDW). Rhein (1991), using freon-11 and freon-12 data, found that even after 1982 there was a transport of GSDW and EBDW into the Norwegian Sea. The average deep volume transport between the Greenland Sea and the Norwegian Sea is estimated at about 1 Sv (Smethie *et al.*, 1988).

The main deep water communication between the Greenland Sea and the Norwegian Sea takes place through a 2000 m deep narrow canyon south in the Jan Mayen Fracture Zone, sometimes referred to as the Jan Mayen Channel (Smethie *et al.*, 1988; Swift and Koltermann, 1988). Sælen (1986) took current measurements during two periods of 7 and 10 months' duration in 1981 and 1983/1984 in this Channel, and on both occasions calculated the transport to be about 0.1 Sv. This indicates that the reduction of GSDW production may be dated to before 1981. However, based on current measurements from 1982, Swift and Koltermann (1988) estimated the transport to be 0.9 Sv. Their current meters were situated east of the narrowest part of the Channel, so they may have overestimated the width of the current. Also, their current meter records are relatively short (about one month).

Based on CTD measurements in 1989 and 1990, Bourke *et al.* (1991) argue that the flow through the Channel had ceased in 1990, and that the newly formed NSDW was only able to penetrate into the Norwegian Sea further north through the Mohn Ridge. However, direct current measurements in several of the channels in the Mohn Ridge during 1988/1989 showed negligible average transports (Foldvik, pers. comm.). The general impression is that the deep flow from the Greenland Sea to the Norwegian Sea continued during the 1980s, but at a low and decreasing rate.

The abyss circulations in the Greenland and Norwegian basins are believed to be cyclonic, steered by the bottom topography (Sælen, 1986). Thus changes in the water characteristics flowing into the Norwegian Sea abyss through the Jan Mayen Fracture Zone will arrive in the bottom layer at "Mike" about one year later, assuming an average current speed of 5 cm/s. The fact that the temperature minimum at 2000 m at "Mike" occurred in 1981, and the temperature has increased since then at this level, without a similar increase at 1500 m, indicates that this change is due to horizontal rather than vertical advection, and that the GSDW production decreased prior to 1980.

The reason for the reduction in the GSDW formation during the last decade is not clear. The cessation seems to coincide with the arrival of the "Great Salinity Anomaly" in the Greenland Sea (Dickson *et al.*, 1988; Schlosser *et al.*, 1991). The reduced salinity will give rise to an increased stability, and therefore at first sight reduce the convection. However, it is demonstrated here (Fig. 4) that when passing the Norwegian Sea the salinity deficit was accompanied by low temperatures, partly compensating for the density reduction by the increased freshwater content. Besides, it has been argued by Aagaard and Carmack (1989) that the most effective way to form GSDW is to cool relative fresh water so that the surface temperature gets colder than the temperature in the water column below. Because of the greater compressibility of cold water, the convection, when started, will have a self-induced amplifier, which will increase the downward penetration of cold water. So it is unclear why the "Great Salinity Anomaly" should reduce the bottom water formation.

Another possible explanation for the reduced bottom water formation is changes in the atmospheric forcing, i.e. increased temperatures. But observations from Jan Mayen and Svalbard show a cooling rather than a warming trend during the 1980s (Nordli, 1990; Hanssen-Bauer *et al.*, 1990). There is a tendency, though, showing the spring temperatures increasing, thus reducing the cooling season. This may be of importance for the bottom water formation, which is believed to take place in the spring after the destratification of the surface water by the winter cooling.

Finally, we would like to mention that deep water temperature fluctuations of the magnitude reported here have been observed earlier: Aagaard (1968) found that the GSDW temperature in 1954 was 0.2°C warmer than in 1910.

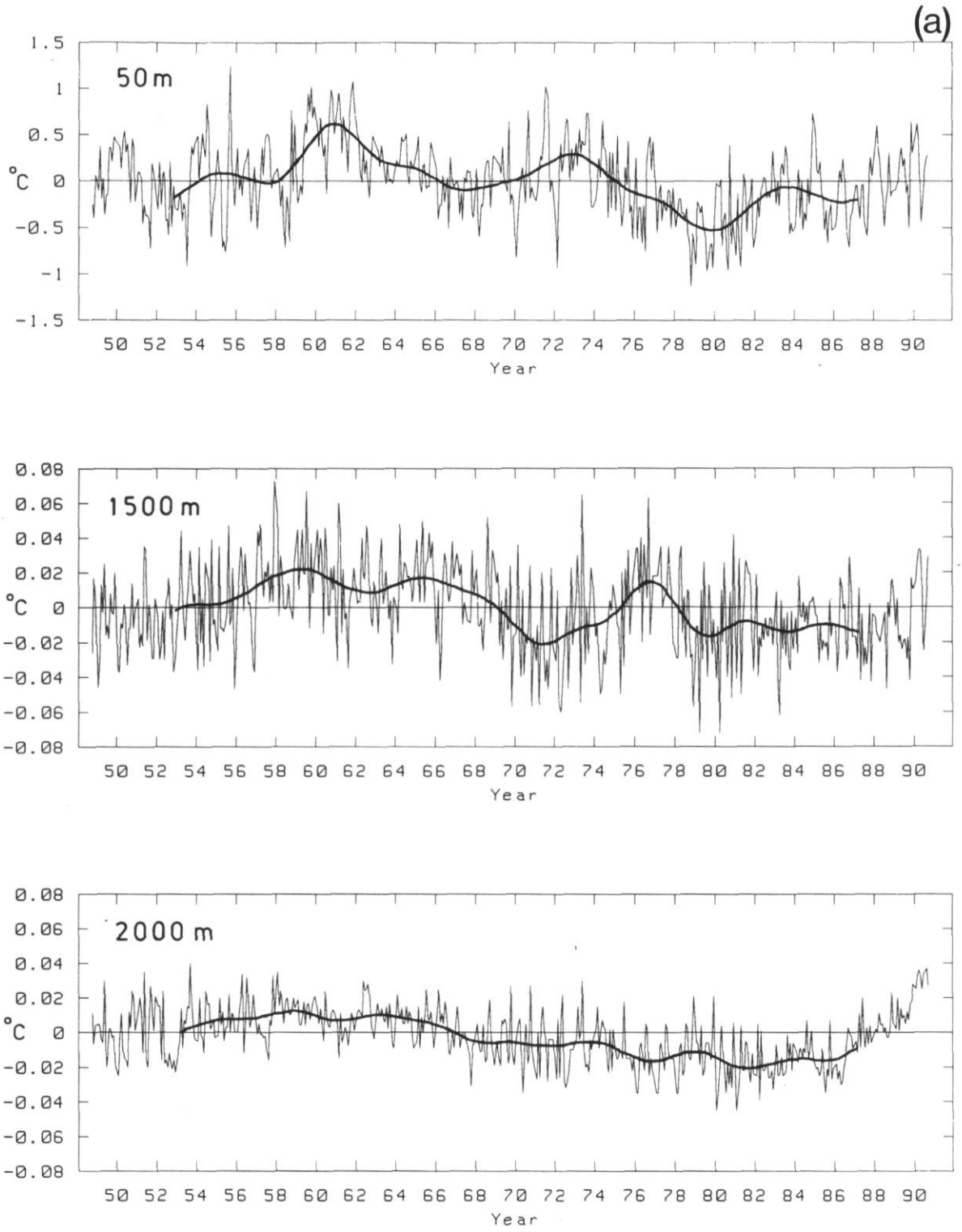
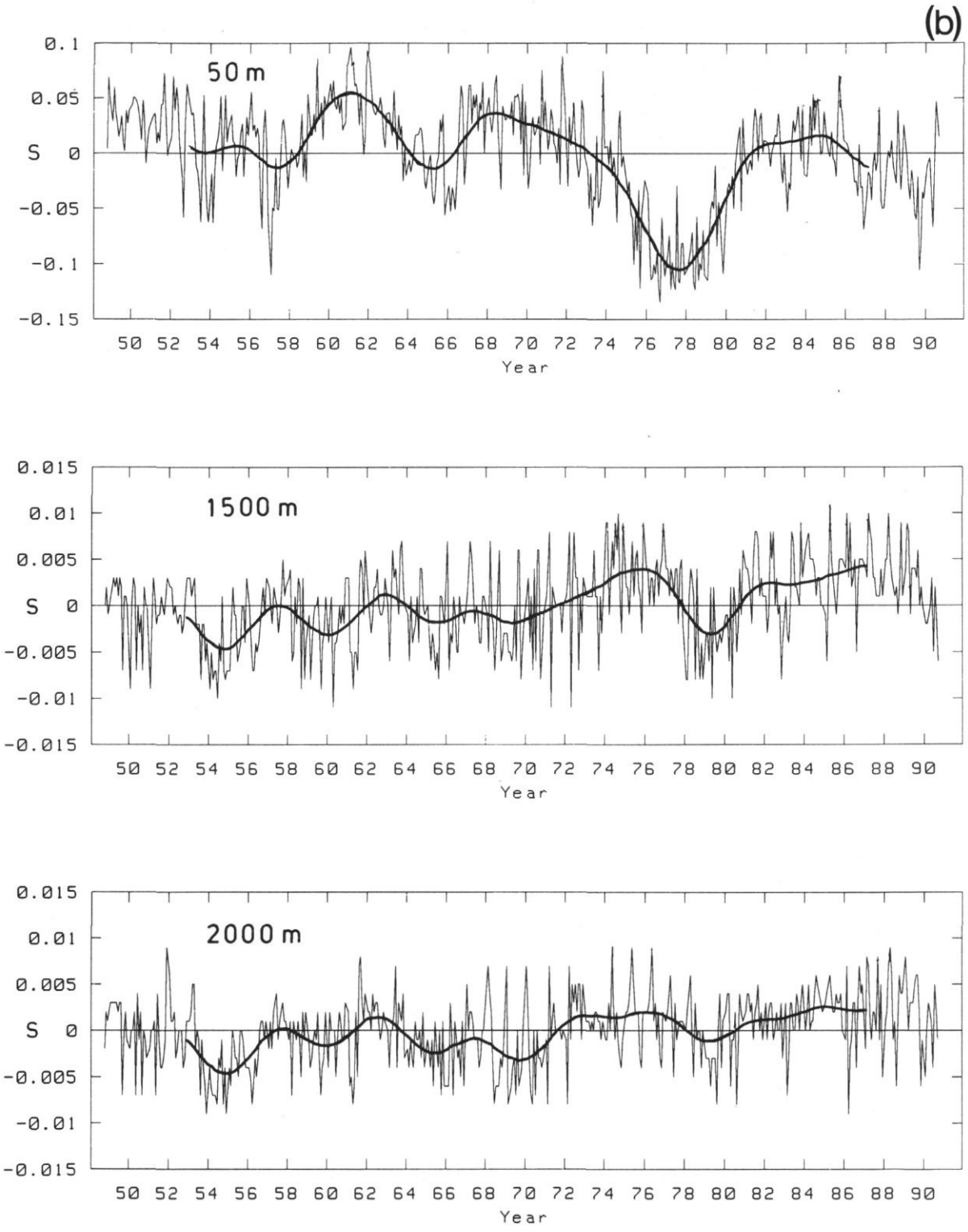


Figure 3. Monthly means of (a) temperature and (b) salinity anomalies at 50 m (representing the surface layer 0–300 m), 1500 m and 2000 m. (Note that the scaling on the ordinates from the surface and deep layers are different.)


(b)
50 m
0.1
0.05
S 0
-0.05
-0.1
-0.15
50 52 54 56 58 60 62 64 66 68 70 72 74 76 78 80 82 84 86 88 90
Year
1500 m
0.015
0.01
0.005
S 0
-0.005
-0.01
-0.015
50 52 54 56 58 60 62 64 66 68 70 72 74 76 78 80 82 84 86 88 90
Year
2000 m
0.015
0.01
0.005
S 0
-0.005
-0.01
-0.015
50 52 54 56 58 60 62 64 66 68 70 72 74 76 78 80 82 84 86 88 90
Year

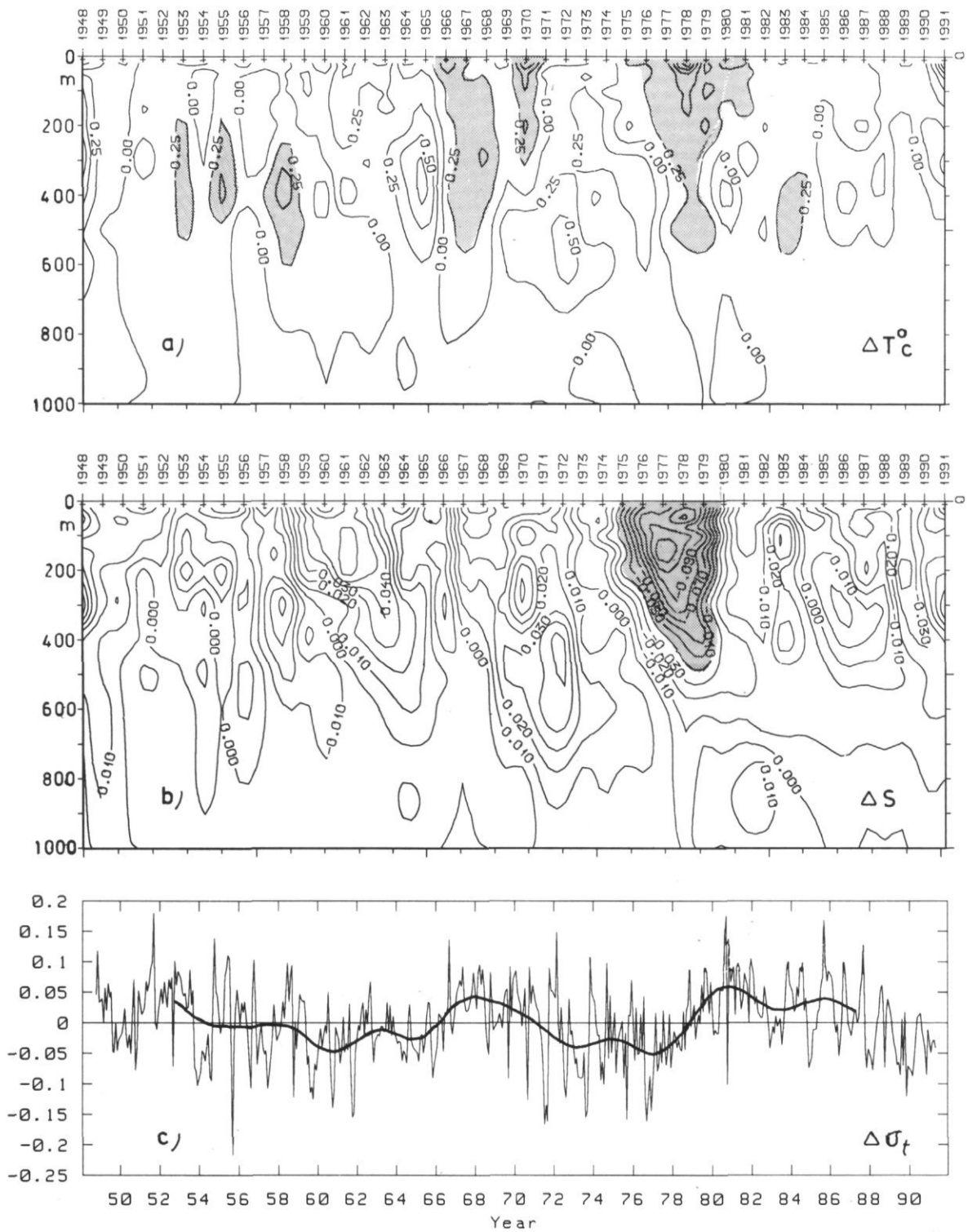


Figure 4. Isopleth diagrams for (a) temperature and (b) salinity anomalies for the upper 1000 m. (Temperature anomalies below −0.25°C and salinity anomalies below −0.05 are hatched.) (c) σ_t anomaly at 50 m depth.

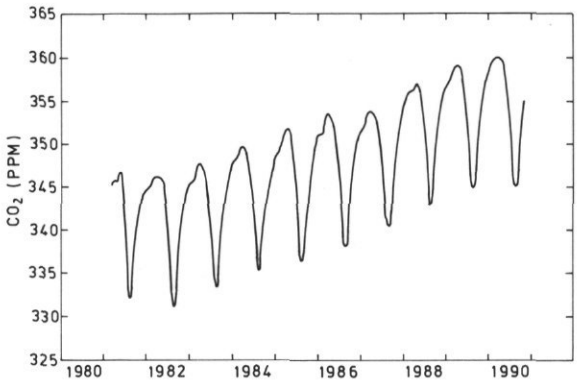


Figure 5. Atmospheric CO_2 concentrations have been observed at Station "Mike" during the last decade by NOAA (Conway *et al.*, 1990).

Future work

"Mike", one of the last weathership stations still running, has a strategic position for climate watch. It is of the greatest importance that this observation series is continued in the same manner to secure homogenous time series.

Other climate relevant parameters are now obtained at "Mike". Since 1980 NOAA has been carrying out registration of greenhouse gases like carbon dioxide (CO_2) (Fig. 5) and methane (CH_4) in the atmosphere (Conway *et al.*, 1990). The trends in the climate gases at "Mike" are comparable to land-based records from the Northern hemisphere (Keeling *et al.*, 1989).

Vertical ocean profiles of carbon (DIC, pCO_2, C^{14},C^{13}) are now obtained regularly in cooperation with the Laboratory for Radioactive Dating (LRD), Trondheim. Tritium and helium profiles will be taken in collaboration with the University of Heidelberg, Germany.

Some of the water samples from the deep ocean have been stored during the years. We hope to be able to construct a 40-year-long time series of radioactive carbon (C^{14}) using the historical water samples from "Mike".

Acknowledgments

We thank the captains and crews on the weathership "Polarfront" for their positive attitude and for the skill they show in carrying out the Nansen Casts in a rough area. They should know that the work they do is of greatest importance in climate research. S.Ø. was supported by the Nordic Council for Physical Oceanography and Ø.G. by the Norwegian Council for Research and the Humanities (NAVF).

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Hydrographic conditions in Icelandic waters, 1980–1989

Svend-Aage Malmberg and Stefán S. Kristmannsson

Malmberg, S-A., and Kristmannsson, S. S. 1992. Hydrographic conditions in Icelandic waters, 1980–1989. – ICES mar. Sci. Symp., 195: 76–92.

The paper follows up data series in Icelandic waters on hydrographic conditions in the East Icelandic Current and in North and South Icelandic waters. In the East Icelandic Current hydrographic conditions were studied in relation to Polar and Arctic water, stratification, and sea-ice conditions. The conditions from 1980 to 1989 in the East Icelandic Current were in general of an Arctic character, with only slight Polar influence in 1982 and 1985 but a major influence in 1988, when sea-ice conditions in North Icelandic waters were relatively severe. The hydrographic observations from North Icelandic waters comprise two seasonal time series from the North Icelandic shelf: indices of warm Atlantic inflow of the Irminger Current and cold Polar water from the East Greenland Current. The Atlantic inflow varied considerably, with a high index in 1980 and 1984–1987. The periods 1981–1983 and 1988–1989 had low indices. The Polar water index was generally low, with few exceptions, notably in 1988. Attempts were made to connect these time series to the wind regime. In 1981–1983 the hydrographic conditions in North Icelandic waters were of an Arctic character, which has been connected with the salinity anomaly in the 1970s in the northern North Atlantic. During these years the Arctic Intermediate water in the Iceland Sea was less saline than during other years until 1989. The hydrographic conditions in South Icelandic waters in the more or less stable warm and saline Atlantic water of the Irminger Current showed a slight decrease in salinity during the years 1985–1988. This anomaly may be connected with outflow of Polar water through the Denmark Strait during the cold ice years in North Icelandic waters in 1975 and 1979 and again with low salinities observed in the Arctic Intermediate water in the Iceland Sea in 1989. In this respect it will in the coming years be of great interest to study the advective nature of the salinity anomaly, especially a possible continuation of the 1975/1979–1985/1988–1989/1993 event in Icelandic waters and its climatological and ecological consequences.

Svend-Aage Malmberg and Stefán S. Kristmannsson: Marine Research Institute, Skúlagata 4, PO Box 1390, Reykjavík, Iceland.

Introduction

Icelandic waters are located at the boundary between warm Atlantic water and cold water from the Arctic, i.e. the oceanic polar front in the northern North Atlantic (Fig. 1). The warm Irminger Current flows northwards and splits into two branches west of Iceland. The western branch meets the East Greenland Current and flows southwards into the Irminger Sea. The eastern branch flows north and eastwards off northwest Iceland into North Icelandic waters (the Iceland Sea), and disperses off the eastern coast of Iceland (Stefánsson, 1962). The cold low-salinity water masses are those of the East Greenland (Polar) Current, flowing southwards through the Denmark Strait, and the East Icelandic Current, flowing southwards north and east of Iceland (Malmberg, 1984a, 1985). This circulation is complicated by north–south fluctuations of the boundary zone from year to year, mainly expressed by the presence of Atlantic, Polar, or Arctic water in North and East Icelandic waters (Fig. 2). These shifts in the location of the oceanic polar front deeply impact the climate and ecology in Iceland as well as the surrounding waters.

This paper presents a follow-up to the data series published previously in *Annales Biologiques* (e.g. Malmberg, 1980, 1983) and elsewhere (Malmberg, 1984a, 1985, 1986, 1988) by describing the hydrographic conditons in Icelandic waters between 1980 and 1989. It includes information on hydrographic conditions (Polar and Arctic) in the East Icelandic Current in relation to sea ice and hydrographic conditions in North Icelandic waters. These data are related to atmospheric conditions over the surrounding sea areas and to salinity anomalies

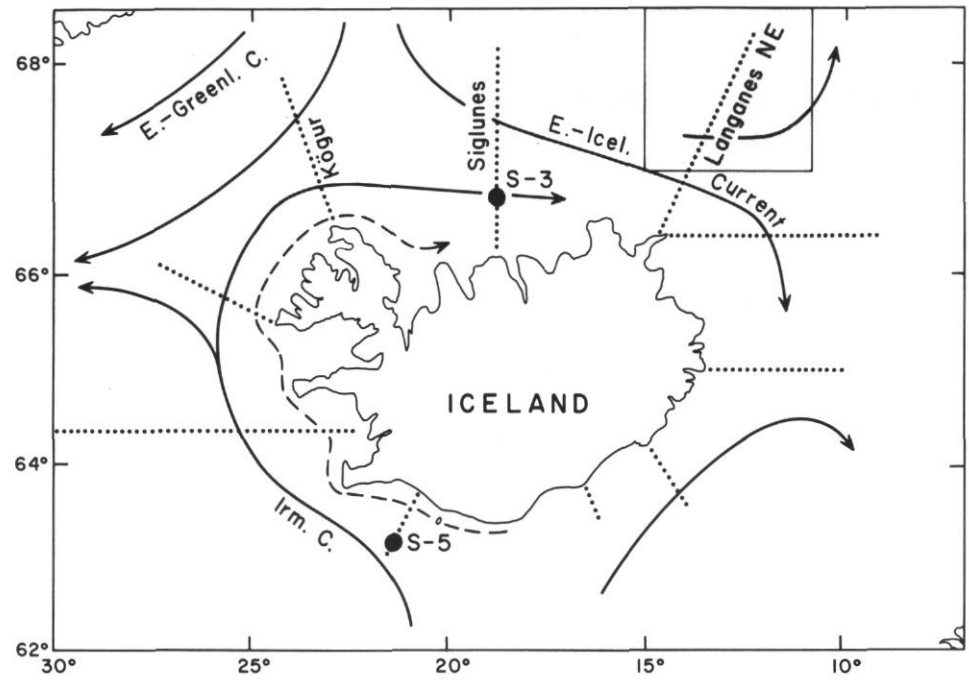


Figure 1. Main ocean currents and location of standard hydrobiological sections in Icelandic waters. Selected areas and stations dealt with in the paper are indicated.

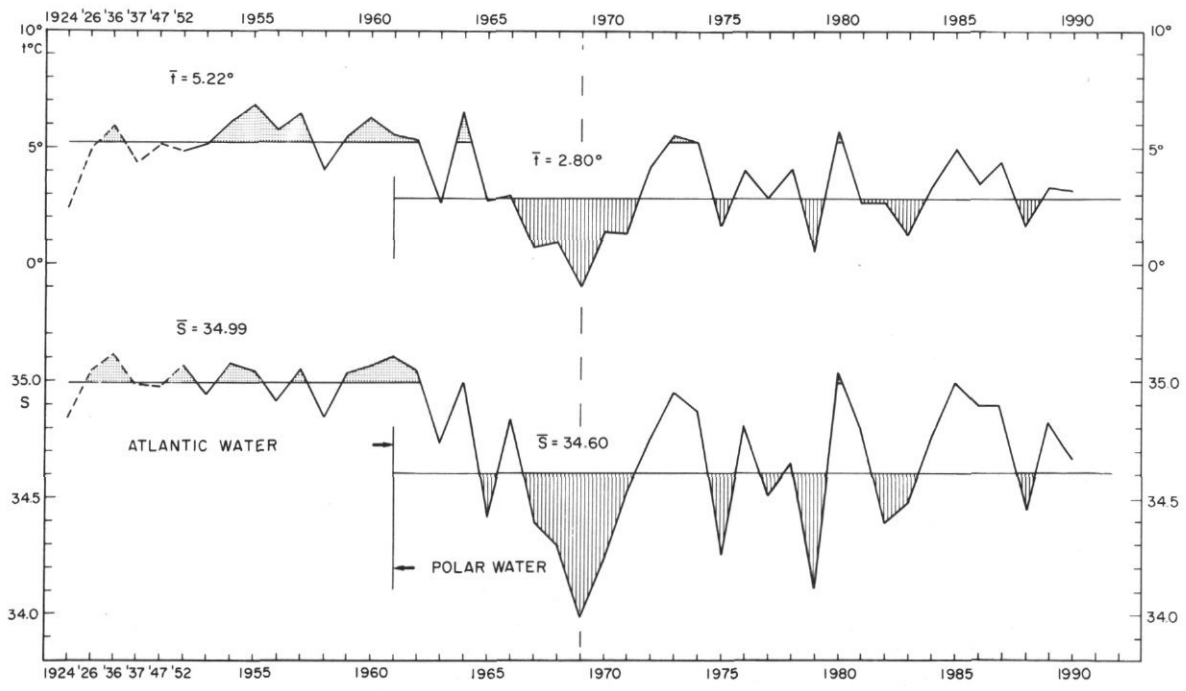


Figure 2. Temperature and salinity at 50 m depth at a hydrographic station in North Icelandic waters (S-3, for location see Fig. 1) in May/June 1924, 1926, 1936, 1937, 1947, and 1952–1990.

Table 1. Mean temperature and salinity within 67–69°N and longitude 11–15°W (shaded area in Fig. 1) in February–March 1981–1990.

	0 m	10 m	25 m	50 m	100 m	200 m	300 m	400 m	500 m
				Temperature (°C)					
1981	−0.58	−0.49	−0.12	−0.55	−0.56	−0.49	−0.03	−0.11	−0.26
1982	−0.42	−0.41	−0.41	−0.42	−0.34	−0.34	−0.28	−0.24	−0.34
1983	−0.32	−0.32	−0.35	−0.35	−0.43	0.00	−0.11	−0.22	−0.32
1984	−0.03	−0.10	−0.01	−0.04	−0.10	−0.17	−0.23	−0.24	−0.31
1985	0.25	0.05	0.00	0.14	0.55	0.30	0.05	−0.21	−0.35
1986	0.28	0.28	0.07	0.26	0.00	0.24	−0.06	−0.22	−0.29
1987	0.33	0.06	0.09	−0.03	0.17	0.42	0.08	−0.18	−0.28
1988	{0.45	0.49	1.07	0.98	0.56	0.09	0.39	−0.08	−0.23}
1989	−0.62	−0.60	−0.57	−0.56	−0.55	−0.15	−0.04	−0.17	−0.22
1990	0.30	0.30	0.30	0.25	0.08	−0.01	−0.07	−0.24	−0.25
				Salinity					
1981	34.78	34.77	34.78	34.78	34.79	34.83	34.88	34.91	34.91
1982	76	76	76	76	77	84	88	91	91
1983	72	73	73	72	73	84	88	91	91
1984	80	80	80	80	81	86	88	90	91
1985	63	63	67	63	84	87	88	90	91
1986	74	74	74	74	75	86	88	90	90
1987	66	67	67	67	72	83	87	90	91
1988	46	46	66	69	72	78	88	90	91
1989	72	70	71	70	71	86	90	91	91
1990	72	71	71	71	72	88	91	91	91

observed in the North Atlantic Current system, including the Irminger Current south of Iceland. The ecological impact of the variable hydrographic conditions will also be outlined.

Since 1970, hydrobiological investigations have been carried out in Icelandic waters on standard sections on a seasonal basis (Fig. 1). Similar investigations were started on a more or less annual basis in the spring of 1948 in connection with herring surveys in North Icelandic waters (Stefánsson, 1962; Jakobsson, 1980; Stefánsson and Jakobsson, 1989).

The study is based on temperature and salinity data from selected areas and stations indicated in Figure 1. All data from the shaded area in the East Icelandic Current were analysed as isopleths of monthly means (Fig. 3, Tables 1, 2), but data from North Icelandic waters and the Irminger Current south of Iceland are simply shown as isopleths from two selected hydrographic stations: S–3 and S–5 (Figs. 5, 11, Tables 3, 6). The isopleths and annual data from selected depths of the isopleths (Figs. 4 and 12) are used for further discussion on stratification and drift-ice conditions in North Icelandic waters (Fig. 4) and on trends or advection of salinity anomalies in the northern North Atlantic in the past, present, and even in the future.

Water masses

The main water masses in Icelandic waters can in general be defined as follows:

(1) Atlantic water of the North Atlantic Drift and the Irminger Current (3–8°C, $S > 34.9$).
(2) Polar water of the East Greenland Current (<0°C, $S < 34.4$).
(3) Arctic/Polar water of the East Icelandic Current (<0–2°C, $S < 34.7$–34.9).
(4) North Icelandic winter water on the North Icelandic shelf (2–3°C, $S = 34.8$–34.9).
(5) Bottom water in deeper layers north and east of Iceland (<0°C, $S \sim 34.89$–34.92).
(6) Arctic Intermediate water below the Polar water in the East Greenland Current (0–2°C, $S \sim 34.95$).
(7) Coastal water on the shelf around Iceland, diluted by runoff (t°C variable, $S < 34.0$).

These water masses are of importance for the ecology in Icelandic waters and the bottom water certainly plays a very significant role in the water mass budget and climatic aspects of the northern North Atlantic (e.g. Swift, 1984).

The East Icelandic Current

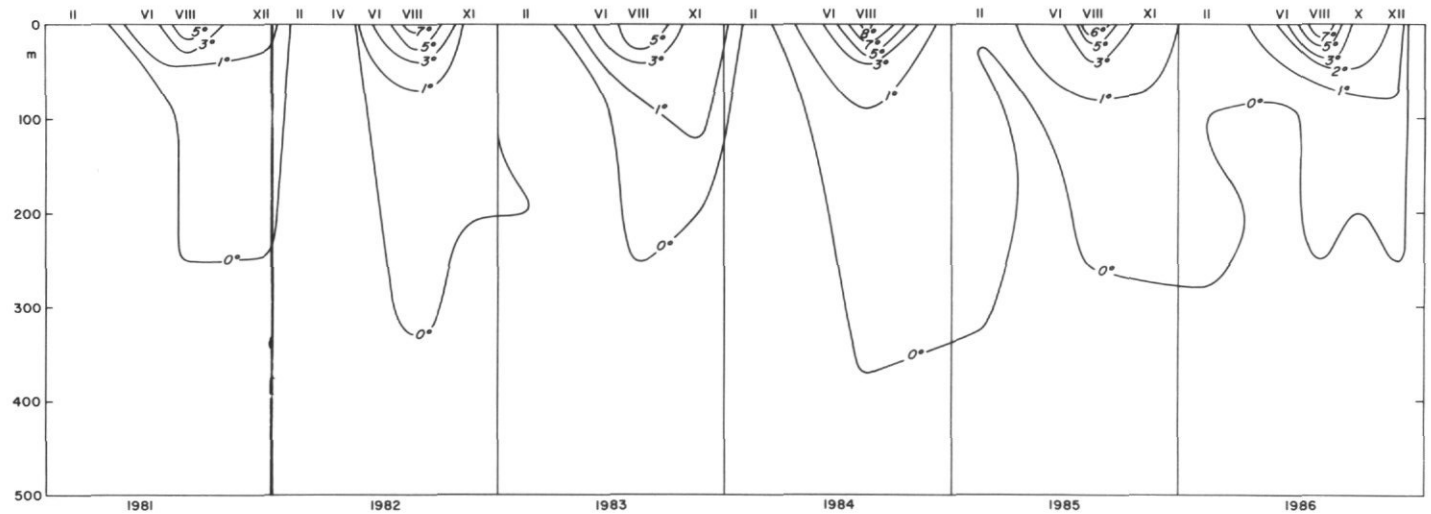
The cold water system in the Iceland Sea north and east of Iceland (Fig. 1), more or less fed by waters from the East Greenland Current and the Irminger Current, is known as the East Icelandic Current (Knudsen, 1898; Kiilerich, 1945; Stefánsson, 1962; Swift and Aagaard, 1981). The East Icelandic Current, an ice-free Arctic current in the period 1948–1963, later advanced south

Table 2. Temperature and salinity deviations within 67–69°N and longitude 11–15°W (shaded area in Fig. 1) in June 1981–1990 from the mean values of the period 1950–1958.

	0 m	10 m	25 m	50 m	100 m	200 m	300 m	400 m	500 m
				Temperature (°C)					
1981	−1.31	−1.49	−1.64	−1.58	−0.65	−0.26	−0.05	0.24	0.22
1982	−1.00	−0.97	−0.95	−0.76	−0.18	0.03	−0.05	0.07	0.14
1983	−0.53	−1.02	−1.26	−0.97	−0.46	−0.22	0.06	0.20	0.16
1984	−0.62	−0.51	−0.56	−0.03	0.12	−0.03	0.02	0.08	0.09
1985	−1.20	−1.45	−0.93	0.05	0.03	−0.12	0.08	0.12	0.11
1986	−1.09	−1.07	−1.04	−0.51	−0.34	−0.12	0.14	0.14	0.10
1987	1.10	0.39	−0.06	0.04	0.04	−0.28	0.02	0.14	0.15
1988	−2.53	−2.49	−2.09	−0.80	−0.36	−0.08	0.18	0.29	0.12
1989	−1.13	−1.31	−1.00	−0.23	−0.34	−0.15	0.18	0.23	0.13
1990	−0.09	−0.59	−0.84	−0.78	−0.24	0.14	0.41	0.27	0.18
				Salinity					
1981	−0.09	−0.09	−0.09	−0.08	−0.06	−0.06	−0.05	−0.01	−0.01
1982	−0.19	−0.19	−0.15	−0.09	−0.06	−0.05	−0.05	−0.03	−0.02
1983	−0.10	−0.11	−0.09	−0.07	−0.07	−0.04	−0.04	−0.02	−0.02
1984	−0.04	−0.04	−0.02	−0.02	−0.02	−0.01	−0.02	−0.02	−0.02
1985	−0.11	−0.11	−0.11	−0.08	−0.05	−0.02	−0.01	−0.01	−0.02
1986	−0.07	−0.07	−0.07	−0.08	−0.07	−0.03	−0.01	−0.01	−0.02
1987	−0.12	−0.12	−0.12	−0.10	−0.08	0.02	0.01	0.00	−0.01
1988	−0.23	−0.23	−0.17	−0.09	−0.08	−0.05	−0.02	−0.01	−0.02
1989	−0.12	−0.11	−0.10	−0.09	−0.07	−0.05	−0.01	0.00	−0.01
1990	−0.11	−0.10	−0.08	−0.09	−0.08	−0.04	−0.01	0.00	−0.01
Means 1950–1958									
T (°C)	3.05	2.96	2.32	0.99	0.12	−0.17	−0.32	−0.36	−0.44
S	34.82	34.82	34.82	34.83	34.85	34.88	34.89	34.91	34.92

Table 3. Temperature and salinity at different depths at station S-3 (66°32′N 18°50′W) in the Siglunes section in May/June 1980–1990.

	0 m	20 m	50 m	75 m	100 m	150 m	200 m	300 m	400 m
				Temperature (°C)					
1980	5.83	5.81	5.67	5.65	5.11	4.26	3.73	1.91	0.32
1981	3.54	3.47	2.59	2.49	2.54	2.27	2.06	1.10	0.69
1982	2.92	2.89	2.63	3.14	3.24	3.16	2.67	0.90	0.34
1983	3.05	2.25	1.23	1.41	1.77	2.39	2.49	1.71	(0.57)
1984	4.84	3.90	3.34	3.22	3.19	3.19	2.65	1.60	0.16
1985	4.89	4.90	4.89	4.80	4.59	4.10	3.59	0.61	0.03
1986	3.84	3.54	3.39	3.34	3.33	3.24	2.63	1.23	0.37
1987	6.08	5.61	4.51	4.51	4.44	3.87	3.24	1.36	0.30
1988	1.99	1.64	1.84	2.84	3.73	2.85	2.30	1.91	1.01
1989	3.45	3.44	3.26	2.48	2.89	2.28	1.64	0.71	0.16
1990	3.73	2.16	2.32	2.53	2.86	2.79	2.19	0.80	−0.05
				Salinity					
1980	35.05	35.06	35.06	35.06	35.02	34.96	34.93	34.88	34.89
1981	34.76	34.76	34.81	34.83	34.87	34.87	34.87	34.84	34.86
1982	34.26	34.26	34.41	34.68	34.75	34.80	34.79	34.75	34.85
1983	34.41	34.41	34.50	34.57	34.71	34.83	34.88	34.86	34.85
1984	34.77	34.81	34.78	34.80	34.83	34.90	34.88	34.87	34.89
1985	34.96	34.98	34.99	35.00	35.00	34.99	34.96	34.90	34.90
1986	34.90	34.89	34.90	34.91	34.92	34.94	34.91	34.88	34.89
1987	34.21	34.79	34.88	34.89	34.90	34.91	34.87	34.83	34.89
1988	33.78	33.96	34.44	34.71	34.86	34.81	34.78	34.80	34.84
1989	34.80	34.80	34.82	34.90	34.81	34.87	34.84	34.85	34.86
1990	34.05	34.24	34.66	34.77	34.84	34.88	34.84	34.77	34.88



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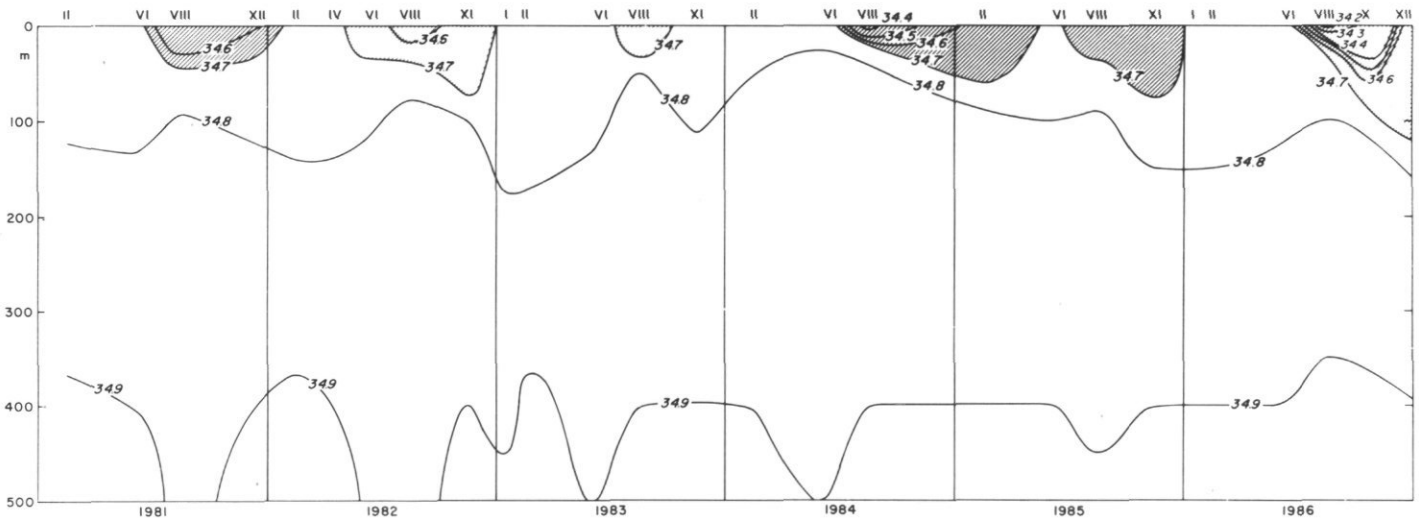

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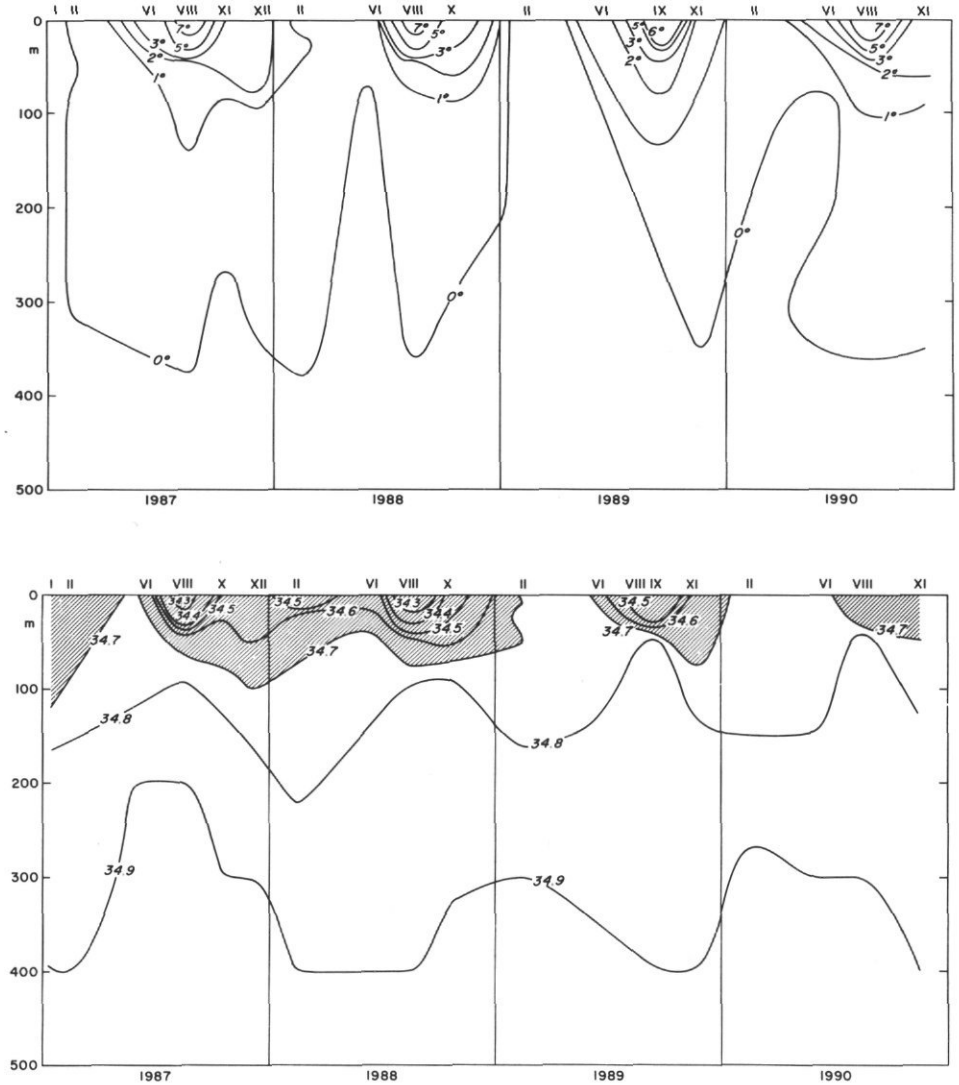

Figure 3. Temperature and salinity isopleths of monthly means for all data collected from the area in the East Icelandic Current shown in Figure 1 during the period 1981–1990.

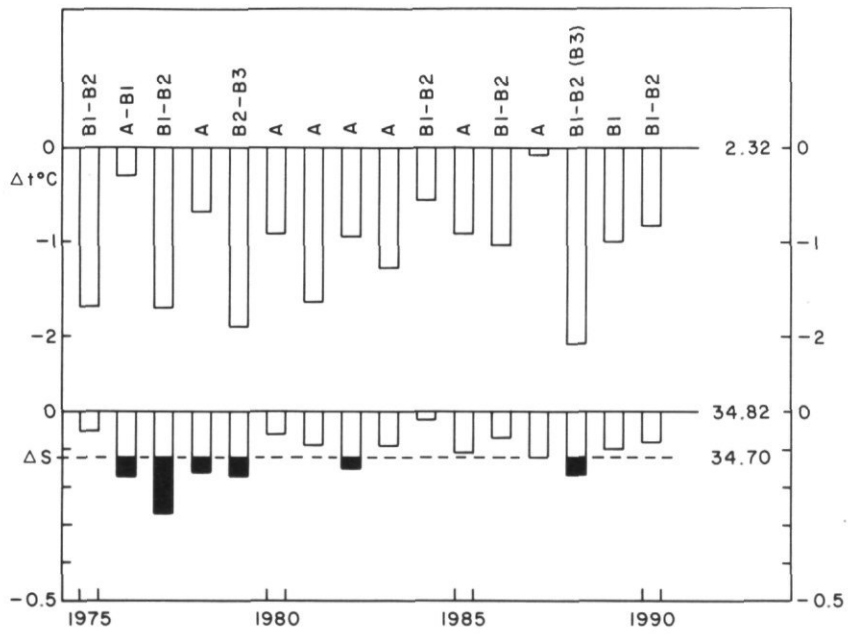

Figure 4. Anomaly of temperature and salinity in spring 1975–1990 at a depth of 25 m in the East Icelandic Current study area shown in Figure 1. The average for the period 1950–1958 is shown as well as a brief classification of ice years in Icelandic waters. Key: A – ice-free; B1 – insignificant ice (NW ice); B2 – moderate ice (N and NW ice); B3 – heavy ice (NW, N, E ice) (from Sigurdsson and Jakobsson, 1991).

and eastwards and developed into a Polar current in 1964–1971, transporting and maintaining drift ice (Malmberg, 1969, 1972).

A correlation was established between the vertical stratification in the East Icelandic Current northeast of Iceland in June and drift-ice conditions in North Icelandic waters in the winter and spring of the same year. A salinity of 34.7 from surface downwards to about 100 m proved the critical value in this area if the surface layers cool down to freezing point, −1.8°C and ice formation is possible. It was suggested that knowledge of stratification in the surface layers of the area north and northeast of Iceland in late winter was of paramount importance in any model for predicting the sea-ice conditions in Icelandic waters in the following spring and summer.

It has been shown that after 1971 the drift-ice conditions in North Icelandic waters and the hydrographic conditons changed back towards former conditions, even returning almost to those prior to 1965 in the years 1972–1974/1975 and 1980 (Dickson *et al.*, 1975; Malmberg, 1984a). In late winter 1975 hydrographic conditons in the East Icelandic Current also indicated an ice-free spring. However, that year more drift-ice and low salinity water were observed in North Icelandic waters than since 1971. In the following years, 1976–1979, both winter and spring conditons generally showed salinities below 34.7 in the study area northeast of Iceland (Malmberg, 1984a). These conditions continued with some drift ice in North Icelandic waters in 1977 and severe ice in 1979 but none in 1978 (Fig. 4). In 1980 the hydrographic conditions in the East Icelandic Current were quite favourable (>34.8) and comparable with findings during the period prior to 1965.

During the years 1981–1990 a salinity as high as 34.8 was observed in the East Icelandic Current in 1984 only (Tables 1, 2; Figs. 3, 4); in the other years it was lower, especially in 1988 when relatively heavy drift-ice conditions were found in North Icelandic waters (Fig. 4; a B3 is missing). The years of low salinity in near surface layers of the East Icelandic Current, 1964–1971, 1976–1979, and 1988 will be discussed later in this paper in relation to a general freshening in the northern North Atlantic and the Nordic Seas in present times, i.e. the so-called salinity anomaly (Dickson *et al.*, 1988). The years 1981–1983 and 1989–1990 seem to be of special interest due to relatively low salinities in intermediate layers (200–300 m, Table 2). This will be discussed in the sections on North and South Icelandic waters.

North Icelandic waters (S-3)

Results of hydrobiological investigations in North Icelandic waters show that in spring Atlantic water ($t > 4°C$, $S > 34.9$) dominated during the periods 1924–1964, 1972–1974, in 1980, and in 1984–1987 (Figs. 2, 5; Table 3). The late 1960s, as well as shorter periods thereafter (1975–1979, 1982, 1988), were more or less characterized by Polar influence, also frequently manifested by the appearance of sea ice (Fig. 4) and by

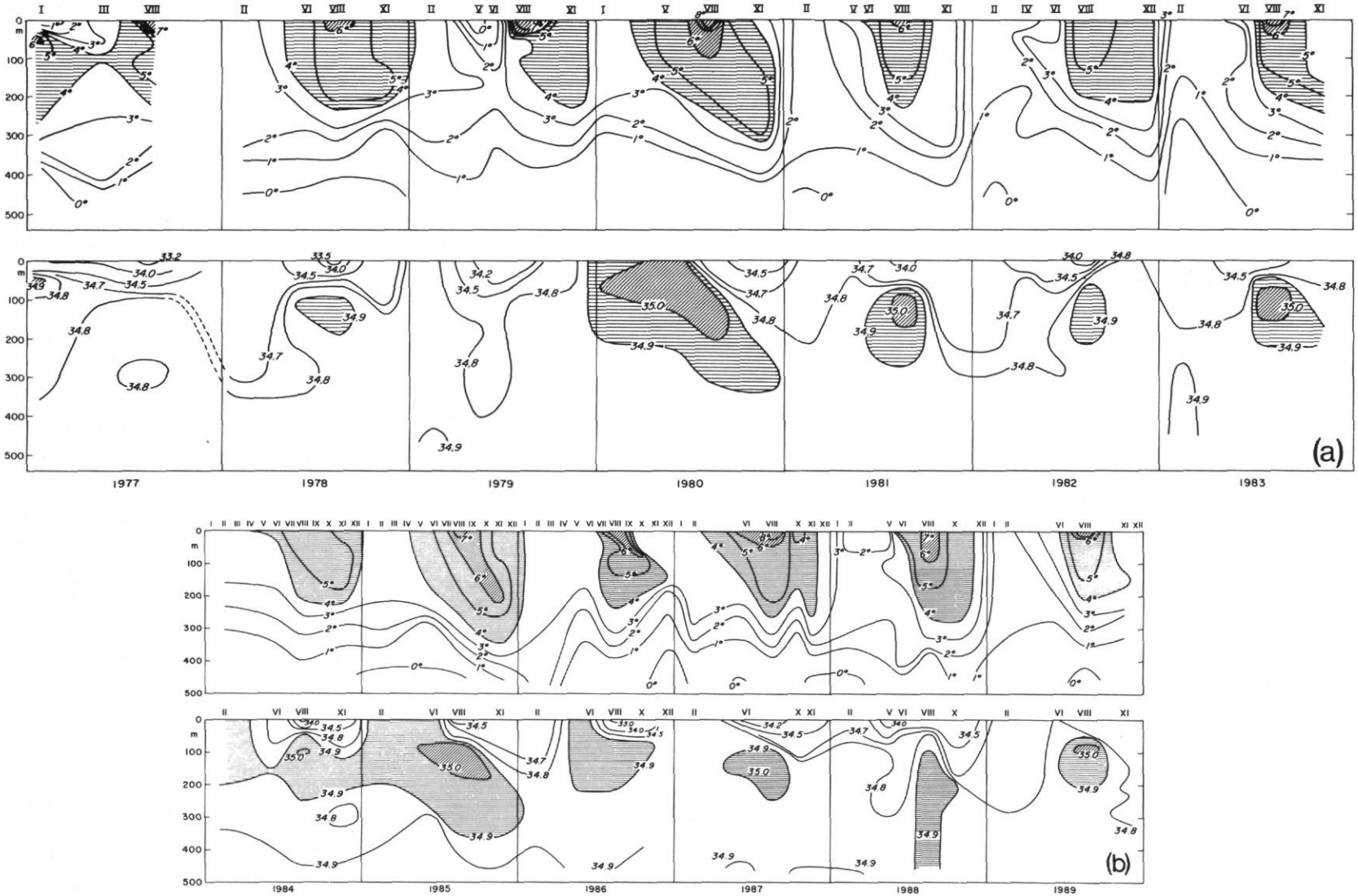

Figure 5. (a) Temperature and salinity isopleths at station S-3 in the Siglunes section, 1977–1983. For location see Figure 1. (From Malmberg, 1984b.) (b) Temperature and salinity isopleths at station S-3 in the Siglunes section, 1984–1989. For location see Figure 1.

biological implications in North Icelandic waters (Malmberg, 1986, 1988; Stefánsson and Jakobsson, 1989). In the years 1981–1983 the hydrobiological conditions in North Icelandic waters were extremely unfavourable and of neither Atlantic nor Polar character, but with a relatively homogeneous water of an Arctic character ($t = 2–3°C$, $S \leq 34.9$; Malmberg, 1986, 1988). These Arctic conditions have been related to the so-called "mid-1970s salinity anomaly" (Dooley *et al.*, 1984; Dickson and Blindheim, 1984; Dickson *et al.*, 1988), which was a decrease in salinity observed in the northern North Atlantic during that time. This decrease was connected with an increased outflow of Polar water from the Arctic (Aagaard and Carmack, 1989) and Nordic Seas during the so-called ice years in Icelandic waters (1965–1971) which was traced from the Denmark Strait (1968) to Labrador (1972) and from there back via the North Atlantic Current System to the coasts of Europe (1976) and into the Nordic Seas (1979–1981). This low salinity water was observed in South Icelandic waters in 1976 (Malmberg, 1977) and in the Iceland Sea north of Iceland in 1981–1983 (Malmberg and Svansson, 1982; Malmberg *et al.*, 1986). In North Icelandic waters this negative salinity anomaly seems to be followed by a relatively weak stratification and hence a homogeneous water mass with poor living conditions for both lower (plankton and larvae) and higher (capelin and cod) living organisms (Malmberg, 1986, 1988).

After the 1981–1983 years of Arctic/Polar conditions in North Icelandic waters the conditions improved again in 1984–1987 to Atlantic conditions (Figs. 2, 5; Table 3), but in 1988–1990 Polar and/or Arctic conditions again took over.

The hydrographic conditions in North Icelandic waters and their variability, which are of vital importance for drift of larvae and recruitment of cod, will be discussed in more detail in relation to seasons and prevailing winds over the surrounding seas in the next section of the paper.

North Atlantic shelf region

The North Icelandic Shelf Region (NISR) at the southern boundary of the Iceland Sea is roughly 65 000 km^2 within the 500 m depth contour. It is defined as the area between the Kögur section in the west and Langanes–NE section in the east (Fig. 1), from nearland and northwards beyond the shelfbreak. Warm Atlantic water enters from the Irminger Current through the Denmark Strait and cold Polar water enters from the East Greenland Current. Both flows are variable in time, as are the hydrographic conditions in the region.

When the warm Atlantic water approaches the west coast of Iceland it has a salinity of about 35.1. By the time it enters the NISR it is a little fresher due to mixing. Therefore, water in the NISR with salinity higher than 34.9 and temperature higher than 3°C is here defined as Atlantic water. Direct current measurements from June 1985 through May 1986 led to an estimated flow into the NISR in the order of 1 Sv (10^6 m^3/s), the mean monthly transports ranging from 0.7 to 2.7 Sv (Kristmannsson *et al.*, 1989). With data from the summer months of 1949–1955 Stefánsson (1962) estimated the net eastward transport of Atlantic water off Kögur to be 0.4 Sv (ranging from 0.1 to 0.6 Sv) using dynamical calculations.

The Polar water originating in the East Greenland Current enters the NISR as a surface layer, most frequently from the northern Denmark Strait area but also from the northwest and north in cold years. The Polar water in the NISR is defined in the top 50 m with salinity below 34.4.

(a) Time series of water masses

Time series from the area have been studied in the past (Malmberg, 1985; Stefánsson, 1969, Fig. 2). They represent the temperature and salinity from selected stations, depths, and seasons. Stefánsson and Guðmundsson (1969) studied variations in the Atlantic water at a station to the northeast of Iceland from August in the period 1934–1965. They found a significant positive correlation between temperature and salinity and southerly winds over Iceland three months earlier.

The time series of water masses studied in this part are formed in the following way: first, data from quarterly hydrographic cruises covering the NISR from 1974–1990 are used. Second, the volume of the water masses in the NISR is integrated and seasonal indices calculated.

The fraction of Atlantic water in the layer between 50 and 300 m was computed. The seasonal mean is shown in Table 4. A yearly cycle was evident with a minimum value of 12% in winter and a maximum value of 37% in summer. Spring and autumn values were 19% and 17%, respectively. This is in agreement with the results of Stefánsson (1962) of increasing inflow of Atlantic water to the NISR during spring, reaching a maximum in summer.

Table 4. Mean Atlantic water component in the 50–300 m layer and mean Polar water component in the 0–50 m layer in the North Icelandic Shelf Region in 1974–1990 as a fraction of total volume.

Season	Atlantic water (50–300 m)	Polar water (0–50 m)
Winter	12	2
Spring	19	20
Summer	37	47
Autumn	17	24

The fraction of the Polar water in the top 50 m in the NISR was also computed. The variability was great with no Polar water in some adjacent seasons but with a large fraction in some years, especially in summer and fall. The mean volume fraction of the Polar water also showed a yearly cycle (Table 4); 2% in winter at minimum but highest 47% in summer. Spring and fall values were 20% and 24%, respectively.

Since there was a marked yearly cycle in these observations the time series of the deviations from the seasonal means have been computed for both water masses (Figs. 6a, b). The Atlantic water revealed four distinct periods in 1980–1990 (Fig. 6a). The unusually warm year of 1980 showed up as a single high above average (1979 was negative). Starting at the beginning of 1981 there was a negatively anomalous period which ended in mid-1983. That roughly marked the beginning of a warm period which ended in the fall of 1987. Continuing through mid-1990 there was a fairly cold period.

The Polar water showed less distinct features (Fig. 6b). The deviation in the Polar water was mostly negative from 1980 through mid-1986, with exceptions in spring 1982 and summer 1984. The years 1986, 1987, and especially 1988 had positive anomalies of Polar water. We note that during 1981–1982 and 1989 both the Atlantic and Polar water components showed negative deviations. This coincided with the presence of Arctic waters in the NISR.

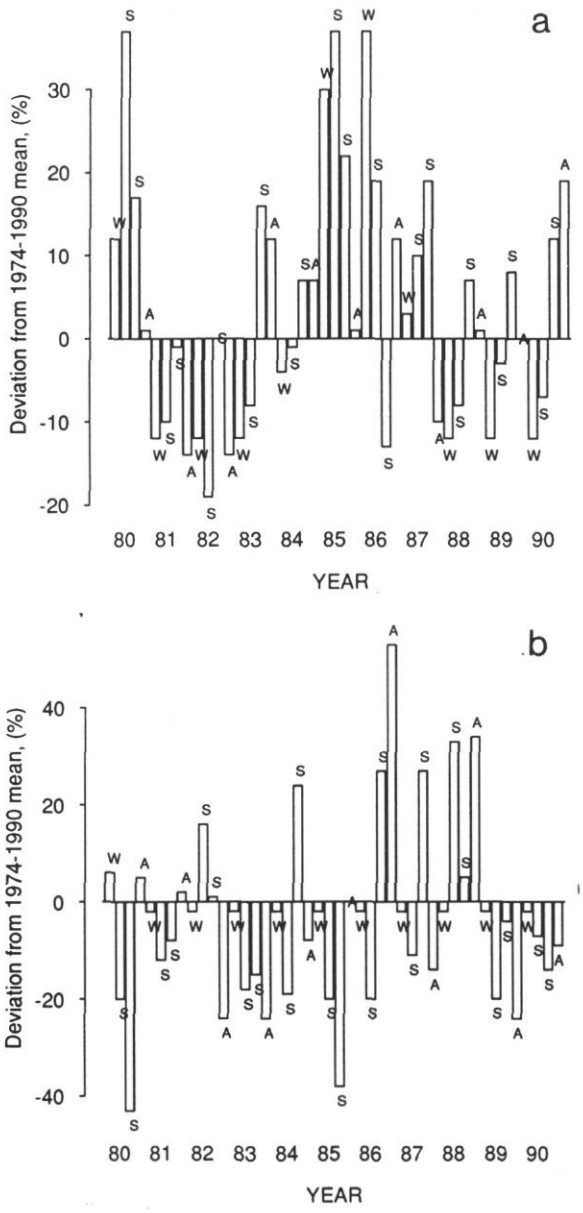

Figure 6. (a) Seasonal index of Atlantic water in the North Icelandic Shelf Region 1980–1990 (W = winter, S = spring, summer, A = autumn). (b) Seasonal index of Polar Water in the North Icelandic Shelf Region 1980–1990 (W = winter, S = spring, summer, A = autumn).

(b) Time series of wind

In this study the Icelandic Low becomes the prime candidate as a possible cause of the variations in water mass distribution in the NISR. Recently, wind data in an accessible format have been available from the hindcast data set of the Norwegian Meteorological Institute in Oslo. The data set covers the North Atlantic, interpolated to a 75 km by 75 km grid at every six hours from 1955. This data set has been studied and used recently in oceanographic problems (Jónsson, 1991).

The variables used here from the hindcast data set were wind stress from the central Denmark Strait and wind stess vorticity (curl) over the Irminger and Iceland Seas. Calculation of the wind stress is based on Large and Pond (1981). Time series were computed from these variables which consist of consecutive three-month averages.

The mean wind stress vorticity over the Irminger Sea had a distinct yearly cycle with values lowest in June but five times higher in December (Fig. 7). The mean wind

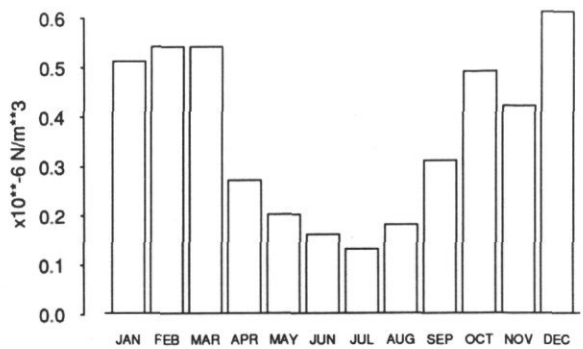

Figure 7. Mean wind stress vorticity over the Irminger Sea, 1974–1988. Computed from the NMI hindcast data set over a 460 000 km^2 area centred at 63°N and 30°W.

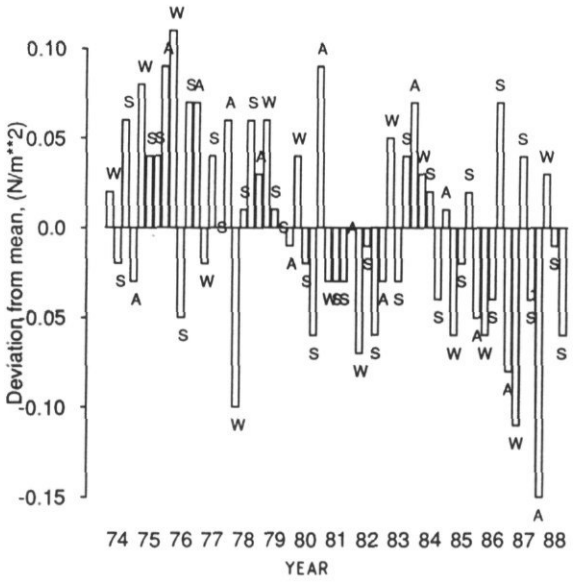

Figure 8. SW/NE seasonal wind stress in central Denmark Strait. Computed from the NMI hindcast data set at 67°N and 25°W (W = winter, S = spring, summer, A = autumn).

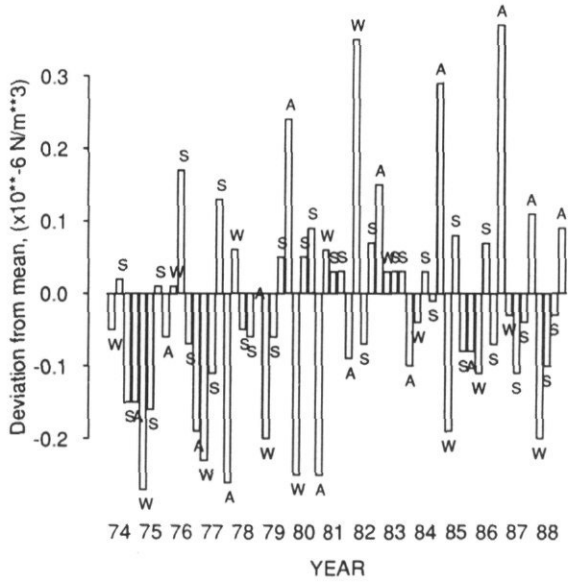

Figure 9. Seasonal wind stress vorticity over the Irminger Sea. Computed from the NMI hindcast data set over a 460 000 km^2 area centred at 63°N and 30°W (W = winter, S = spring, summer, A = autumn).

stress vorticity over the Iceland Sea also had a similar yearly cycle but the values were lower and not as coherent. However, the variance of the wind stress vorticity was higher in the Iceland Sea than the Irminger Sea. Time series of the deviations in wind stress from the seasonal mean were computed in the Denmark Strait (Fig. 8). Time series of the deviations in wind stress vorticity from seasonal mean were also computed for the Irminger and Iceland Seas (Figs. 9, 10). These time series were highly variable. The wind stress vorticity over the Irminger Sea showed that the Icelandic Low was generally less intense before 1980 in that area. This was emphasized over the Iceland sSa, where the wind stress vorticity deviation was almost continuously negative from 1974 to 1982 but continuously positive from 1984 to mid-1987. Wind stress in the Denmark Strait had a decreasing trend from 1974 to 1988, indicating a strengthening between the Greenland High and Icelandic Low during that period.

These three sets of time series were used for comparisons with the Atlantic and Polar water in the NISR. Significant linear correlations were found in the following (Table 5):

(1) A positive correlation between the wind stress in the Denmark Strait (southwesterlies) and the Polar water in the NISR about two months later. However, this correlation was only marginal and not significant with a lag of three months.
(2) A positive correlation between the wind stress vorticity over the Irminger Sea, and even stronger over the Iceland Sea, and the Atlantic water in the NISR 6–9 months later.
(3) A negative correlation between wind stress vorticity over the Iceland Sea and in particular the Irminger Sea and the Polar water in the NISR nine months later.

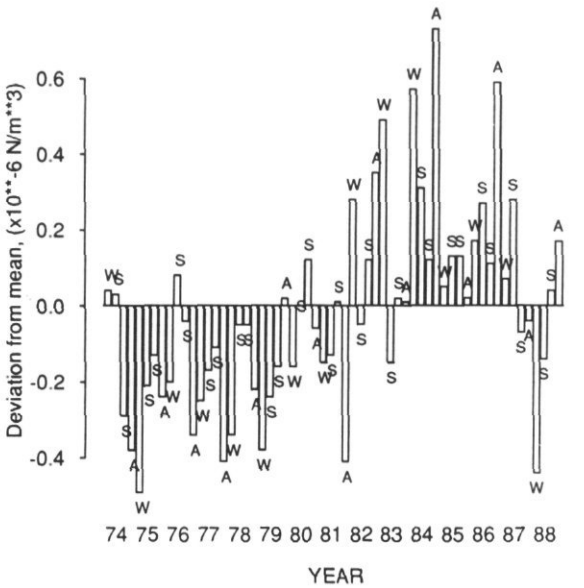

Figure 10. Seasonal wind stress vorticity over the Iceland Sea. Computed from the NMI hindcast data set over a 250 000 km^2 area centred at 68°N and 18°W (W = winter, S = spring, summer, A = autumn).

Table 5. Correlation coefficients between seasonally adjusted wind variables and watermass indices (WSVIR = wind stress vorticity (curl) over the Irminger Sea, WSVIS = wind stress vorticity over the Iceland Sea, WSDS = wind stress in the Denmark Strait, AW = Atlantic water, PW = Polar water); positive lag: wind leads water; length of time series = 60.

Lag (months)	WSVIR/AW	WSVIR/PW	WSVIS/AW	WSVIS/PW	Lag (months)	WSDS/PW
−12	−0.08	0.02	0.10	−0.17	−13	−0.23
−9	0.12	−0.07	0.13	−0.11	−10	0.17
−6	−0.05	0.03	0.26	−0.10	−7	0.07
−3	−0.10	0.35	0.21	−0.02	−4	−0.25
0	−0.02	−0.09	0.26	−0.15	−1	0.10
3	−0.02	−0.27	0.26	−0.21	2	0.28
6	0.27	−0.10	0.44	−0.22	5	−0.17
9	0.27	−0.39	0.44	−0.31	8	0.13
12	0.09	−0.11	0.26	−0.04	11	0.07

The first result is probably easiest to interpret. It simply states that when the wind was from the southwest for some time, Polar water drifted away from the East Greenland Current and into the NISR for up to two months later. On the other hand, northeasterlies pressed that current closer to the coast of East Greenland. In Table 5 there is also the implication that once Polar water was in the NISR the result was increased wind from the northeast four months later.

The conclusion from the second result is that a strong low over the Irminger and Iceland Seas resulted in an increased inflow of Atlantic water to the NISR 6–9 months later. This was particularly true for fall/winter weather and spring/summer Atlantic water. In terms of the Icelandic Low this means that if the storm tracks lay to the west of Iceland and the cyclones passed through the Denmark Strait, rather than to the south and east of Iceland, the result was increasing warm Atlantic water in the NISR 6–9 months later. The Iceland Sea wind stress vorticity and Atlantic water correlation was even marginally significant at the 95% level with a time lag of 0–3 months. This can be attributed to a reduced Greenland High, and thus increased southerlies resulted in increased Atlantic water at the same time and extending nine months into the future.

At least two possible processes can be at work. One is increased wind stress vorticity input in the Irminger Sea driving the water northward onto the shelf off Iceland and eastward through the Denmark Strait. An alternative explanation is that strong storm activity over the Iceland Sea in winter, possibly resulting in anomalous amounts of intermediate water formation north of Iceland, would somehow, because of continuity, lead to increasing inflow of Atlantic water through the Denmark Strait in the following spring.

The third result implies the following. Since Polar water is at its seasonal maximum in summer, a strong Icelandic Low in the fall results in decreased amounts of Polar water the next summer in the NISR. This implies that when the storm tracks were over the Iceland Sea in fall and the associated cyclonic activity, thus inducing more vertical mixing and cooling of Arctic waters in the Nordic Seas than usual, the result was minimum Polar water in the following summer. Also implied in Table 5 is a swifter response to a deeper Icelandic Low over the Irminger Sea resulting in decreasing Polar water in three months. Further, with Polar water already in the NISR the results imply that the water influenced the weather in that case. In three months the presence of Polar water resulted in the Greenland High extending further to the east and thus reducing the wind stress vorticity over the Irminger Sea.

South Icelandic waters (S-5)

The Selvogsbanki section crosses the South Icelandic shelf on the main spawning field of cod (Fig. 1). The section includes the coastal water with its dilution by runoff as well as the Atlantic water of the Irminger Current. The local variations in the coastal water seem to be of importance for the spawning of cod (Ólafsson, 1985; Jónsson and Friðgeirsson, 1986; Thórdardóttir, 1986). The slight variations observed in the Atlantic water are connected with large-scale features in the open ocean (Neumann, 1940; Smed, 1975; Rodewald, 1972).

The "1970s salinity anomaly" in the northern North Atlantic mentioned above was observed south of Iceland in 1976 or even during the years 1974–1978 (Malmberg, 1984b and Figs. 11, 12). It was related to the ice-years north of Iceland in 1965–1971 about eight years earlier. Later on in 1981–1983, it was observed in North Icelandic waters, as mentioned above, or about five years later than south of Iceland.

The investigations in the international Greenland Sea Project (1987–1991) in the western part of the Iceland Sea have shown (Malmberg *et al.*, 1990) that salinity in the Arctic Intermediate water had decreased slightly in 1989–1990 compared with 1987–1988 (34.92 vs 34.95). The "low" salinity may possibly be traced back to South Icelandic waters in the years 1985–1988 when a slight salinity minimum was observed (Figs. 11, 12; Table 6) and again to the polar water years in North Icelandic waters and in the East Icelandic Current observed in 1975–1979 (Figs. 2, 4, 12). The low salinity in the Arctic Intermediate water north of Iceland observed in 1989/

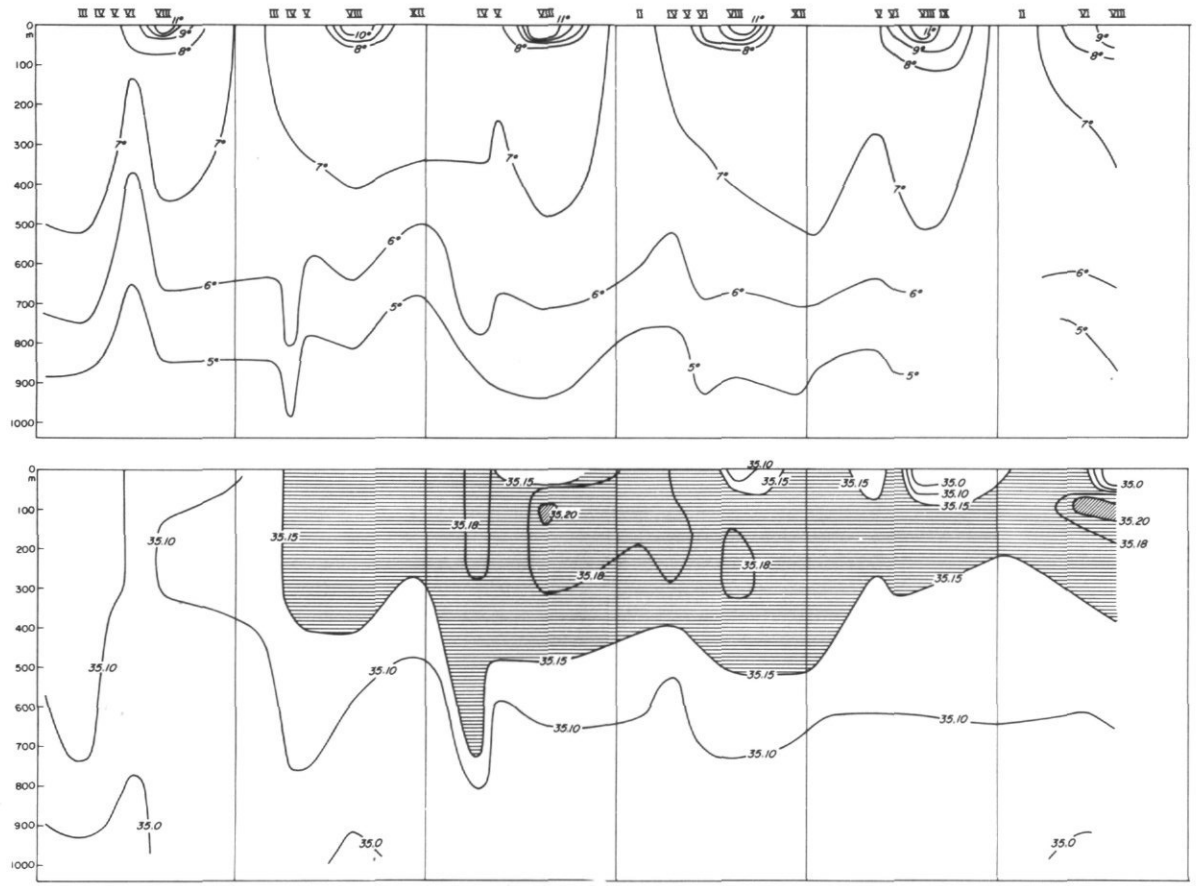

Figure 11a. Temperature and salinity isopleths at station S-5 in the Selvogsbanki section, 1978–1983. For location see Figure 1. (From Malmberg, 1985.)

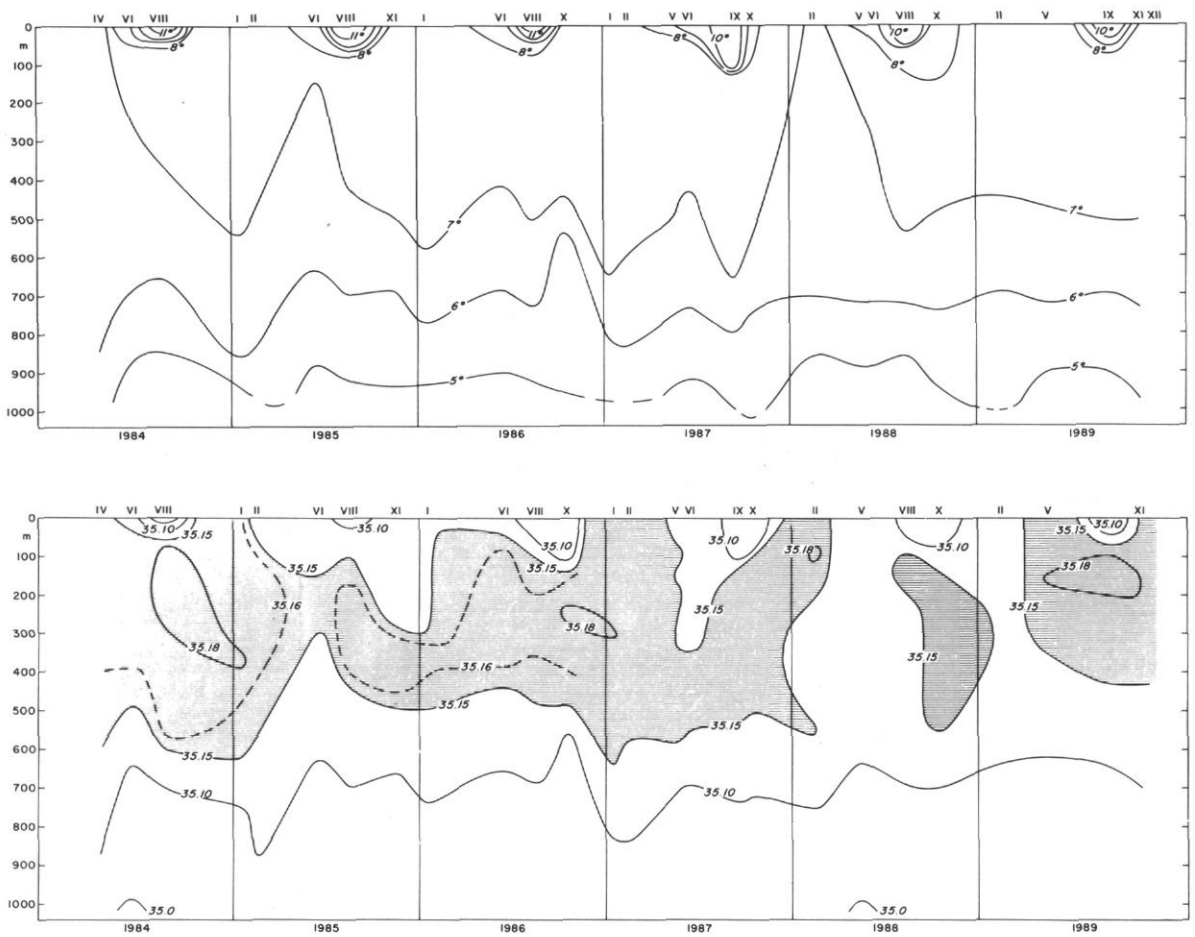

Figure 11b. Temperature and salinity isopleths at station S-5 in the Selvogsbanki section, 1984–1989. For location see Figure 1.

Table 6. Temperature and salinity at different depths at station S-5 (63°00′N 21°28′W) in the Selvogsbanki section in May/June 1981–1989.

Depth (m)	1981	1982	1983	1984	1985	1986	1987	1988	1989
				Temperature (°C)					
0	8.10	7.98	8.49	9.07	8.80	8.52	8.13	8.27	7.60
10	7.93	7.93	8.48	9.10	8.76	8.51	8.12	8.14	7.57
20	7.76	7.95	8.47	9.09	8.57	8.49	8.07	8.04	7.56
30	–	7.94	8.42	8.94	8.52	8.49	8.00	7.99	7.54
50	7.40	7.82	8.32	7.92	8.35	8.09	7.48	7.85	7.58
75	7.32	7.33	7.68	7.52	7.51	7.69	7.41	7.68	7.57
100	7.27	7.19	7.56	7.35	7.20	7.53	7.34	7.66	7.54
150	7.22	7.11	7.19	7.18	7.00	7.38	7.34	7.08	7.44
200	7.15	7.10	7.06	7.05	6.89	7.28	7.32	7.07	7.36
300	6.95	6.99	6.90	6.90	6.79	7.14	7.19	6.96	7.21
400	6.69	6.83	6.70	6.80	6.59	7.02	7.12	6.90	7.13
500	6.46	6.62	6.50	6.64	6.45	6.82	7.01	6.79	6.94
600	6.04	6.20	6.22	6.39	6.15	6.45	6.82	6.60	6.63
800	5.05	5.11	4.90	5.52	5.25	5.34	(5.58)	5.55	5.60
1000	4.40	–	(4.11)	(4.55)	–	(4.73)	(4.63)	4.46	(4.77)
				Salinity					
0	35.15	35.13	35.16	35.13	35.13	35.12	35.14	35.12	35.16
10	35.15	35.13	35.16	35.13	35.12	35.12	35.16	35.11	35.16
20	35.16	35.14	35.16	35.13	35.13	35.12	35.15	35.12	35.16
30	35.16	35.13	35.16	35.13	35.12	35.12	35.15	35.11	35.16
50	35.17	35.14	35.17	35.18	35.13	35.15	35.14	35.11	35.16
75	35.17	35.15	35.20	35.17	35.14	35.16	35.15	35.11	35.16
100	35.18	35.15	35.20	35.17	35.14	35.16	35.16	35.11	35.16
150	35.18	35.15	35.18	35.16	35.15	35.16	35.15	35.11	35.18
200	35.18	35.15	35.17	35.16	35.15	35.16	35.16	35.13	35.16
300	35.17	35.15	35.15	35.16	35.14	35.15	35.15	35.13	35.16
400	35.15	35.14	35.14	35.15	35.14	35.15	35.15	35.13	35.14
500	35.14	35.12	35.12	35.15	35.13	35.13	35.15	35.12	35.13
600	35.11	35.10	35.10	35.12	35.10	35.12	35.15	35.11	35.10
800	35.04	35.04	35.01	35.05	35.05	35.06	35.07	35.05	35.05
1000	30.03	–	(34.98)	(35.00)	–	(35.03)	(35.02)	35.00	(35.01)

1990 may thus possibly last a few years more, or up to 1992. Furthermore, if the low saline water found in the East Icelandic Current and in North Icelandic waters in 1988 follows its previous pathway from North to South Icelandic waters, it may be observed again south of Iceland around 1996.

Conclusions

Signs of the hydrographic conditions in the Iceland Sea and their far-reaching influences may be seen in the conditions of the Icelandic capelin stock at their lowest during the present years around 1990 as in 1982 (Fig. 13; Vilhjálmsson, pers. comm.). Feeding conditions and migration paths of the capelin may have been threatened because of weak stratification in the Iceland Sea. Furthermore, this again is reflected in the weight of cod in Icelandic waters as shown for five-year-old cod in Figure 13 (Anon., 1991), since cod feeds to a high degree on capelin (Pálsson, 1983).

It will also be interesting to follow the 1988 salinity minimum found in the East Icelandic Current (Figs. 3, 4, 12; Tables 1, 2) and in North Icelandic waters (Figs. 2, 5, 12; Table 3) to see if it will be found south of Iceland around 1996 and again north of Iceland around the year 2000, repeating the eight- and five-year cycles from the 1970s. The ecological impact of these conditions in North Icelandic waters has been stated (Fig. 13; Malmberg, 1986, 1988). But also, the negative overall role of a freshening in the Nordic Seas as regards ventilation or deep water formation is noted, which again may have an impact on the flow of Atlantic water northwards, and hence climate in these northern regions and even western Europe.

Finally the hydrographic time series from Icelandic waters described in this paper are not separate oceanographic problems but are closely related to processes such as stratification due to salinity and advection over large ocean areas in connection with air–sea interaction. Thus these time series may be of use for modelling to understand better the physical-biological processes underlying the ocean and biological variability, and help us to look forward with some prediction in mind. It is the belief of the authors that a major factor concerning variations in size and behaviour of fish stocks, at least along the oceanic polar front, is the maritime climate. It influences the living conditions including spawning, feeding, migration, recruitment, growth, and maturity.

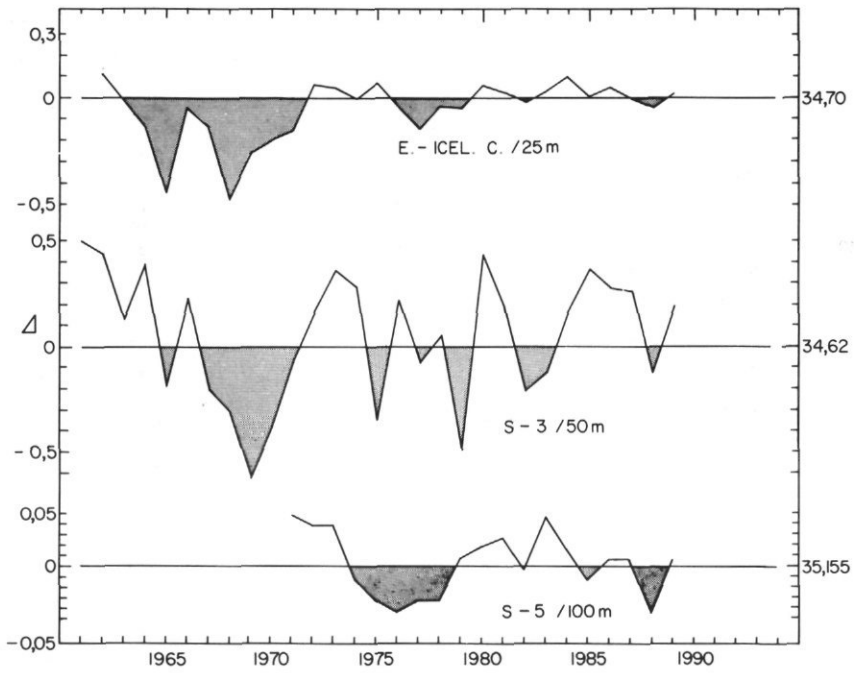

Figure 12. Salinity deviations in spring at 25 m in the East Icelandic Current, 1962–1989; at 50 m in North Icelandic waters (S-3), 1961–1989; and at 100 m in the Irminger Current south of Iceland (S-5), 1971–1989.

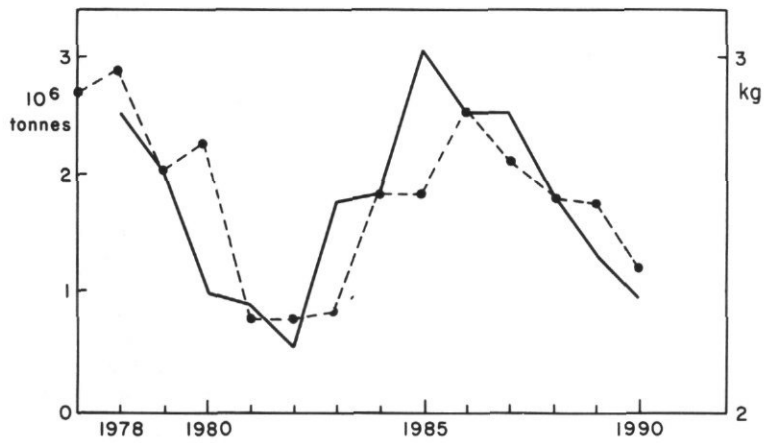

Figure 13. The abundance of the Icelandic capelin stock in 1978–1990 (Hjálmar Vilhjálmsson, pers. comm., solid line) and weight of five-year-old cod in Icelandic waters in 1977–1990 (Anon., 1991, broken line).

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ICES mar. Sci. Symp., 195: 93–102. 1992

The 1980s in the context of climatic changes in the North Atlantic region

S. N. Rodionov and A. S. Krovnin

Rodionov, S. N., and Krovnin, A. S. 1992. The 1980s in the context of climatic changes in the North Atlantic region. – ICES mar. Sci. Symp., 195: 93–102.

The 1980s may be considered as having been very warm years in terms of global surface air temperature. The question whether this warming has been reflected in atmospheric and oceanic processes in the North Atlantic and the ICES area in particular is discussed. Analysis of decadal sea-level pressure and 500 hPa heights anomaly maps shows that both the 1980s and the 1970s were characterized by strong westerlies over the North Atlantic region; this was in contrast to the 1960s, when anomalous easterly winds prevailed. However, an enhanced zonal circulation did not result in a clear tendency toward warming in the Northeast Atlantic in the 1980s, as one could expect, though both air and sea surface temperature were abnormally high in 1983 and 1989. In general, the 1980–1989 decade was the coldest in the North Atlantic since the end of the last century. For the last 30 years, it is possible to distinguish three climatic states of the North Atlantic Oscillation (NAO) which resulted in three well-pronounced climatic regimes in the Northeast Atlantic: 1963–1969, 1970–1976, and 1977–1982. The rest of the 1980s cannot be attributed to any particular regime. During these years both record low and high values of the index of the NAO were registered. Increased climatic variability is the most prominent feature of the 1980s.

S. N. Rodionov and A. S. Krovnin: All-Union Research Institute of Marine Fisheries and Oceanography (VNIRO), 17a Verkhnyaya Krasnoselskaya, Moscow, 107140, Russia.

Introduction

The 1980s were very warm years in terms of global air surface temperature. As some scientists (Budyko and Groisman, 1989; Hansen *et al.*, 1988) assume, this may be considered as a signal that global warming due to an increasing level of greenhouse gases in the atmosphere is under way. Though the warming of the 1980s was global, its extent differed between the Northern and Southern Hemispheres. Whereas air surface temperature was higher in the Southern Hemisphere in the last decade than in any previous decade (Anon., 1988), the 1980s were on average 0.02° and 0.03°C colder than the 1930s and 1940s respectively in the Northern Hemisphere, taking a median as an estimate of the mean decadal value (Fig. 1). However, four of the five warmest years in the Northern Hemisphere were observed during the 1980s. In descending order these were 1988, 1981, 1983, 1989, and 1987.

The main objective of the present paper is to consider the development of large-scale processes in the atmosphere and ocean in the North Atlantic against a background of global warming. As the North Atlantic Oscil-

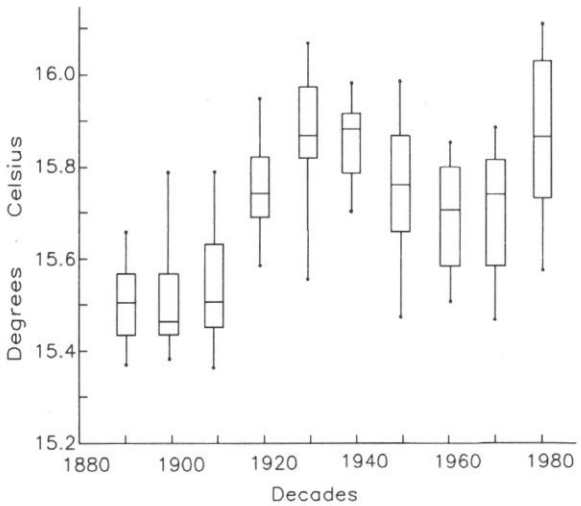

Figure 1. Changes in decadal air surface temperature in the Northern Hemisphere. Horizontal lines inside boxes indicate median. Length of the box is difference between upper and lower quartiles. Dots are minimum and maximum values of temperature within each decade. Based on annual air temperature data set obtained by Gruza *et al.* (1990).

lation (NAO) is the dominant feature of the climatic system in the region under consideration, our main attention is directed to those peculiarities of the NAO during the 1980s which are of greatest importance for the ICES area.

Data and methods

As a basis for the study, monthly mean surface air temperatures (SAT), sea surface temperatures (SST), sea level pressures (SLP), and 500 hPa heights (H_{500}) from the USSR Hydrometeorological Center data archives were used. Data on SAT, SLP, and H_{500} were on the 5° latitude by 10° longitude grid and those on SST on the 5° × 5° grid. Main attention is given to the winter season when the NAO is most pronounced, as demonstrated by Figure 2. In this figure, the one-point correlation maps, showing the correlation coefficients between air temperature at grid point 70°N 30°E and air temperature at every grid point in the North Atlantic region, are presented. As seen from the figure, a classical four-pole pattern in the SAT field with air temperature anomalies of the same sign in the northeastern and southwestern North Atlantic and opposite anomalies in its northwestern and southeastern parts is most prominent only in the winter season. So, all computations except those specially mentioned were made for winter. Winter represents data from the months of December, January, and February. For example, winter 1981 includes December 1980 along with January and February 1981.

The index of the NAO was constructed as the difference of SAT between two grid points located close to the stations used by van Loon and Rogers (1978) – the first near Oslo, Norway (60°N 10°E), the second near Jakobshavn, Greenland (70°N 50°W). To make variations in these two points comparable, normalized mean winter anomalies were used.

Data on central pressures in the Icelandic Low and Azores High were taken from mean winter maps. To determine the position of these centres of action, mean

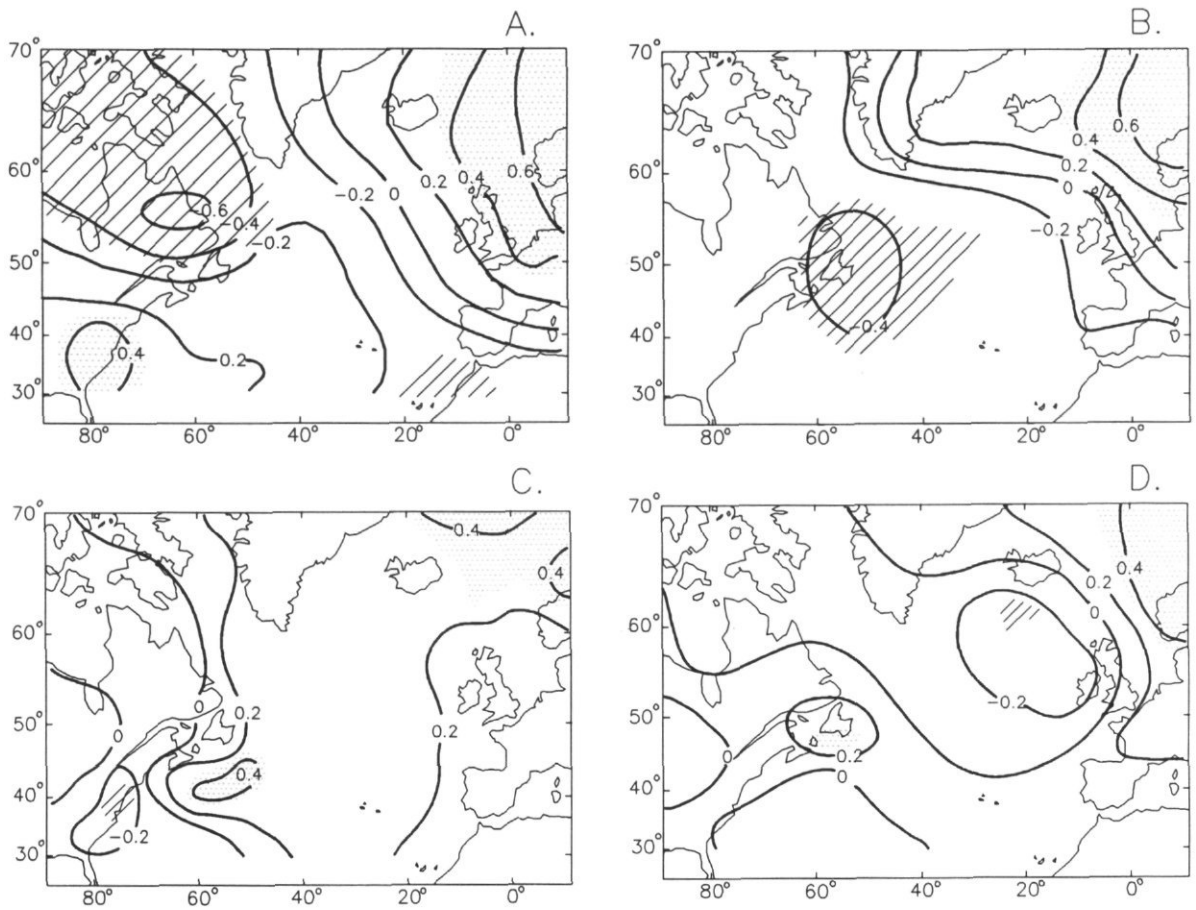

Figure 2. One-point correlation maps showing correlation coefficients between mean seasonal surface air temperature (SAT) at grid point 70°N 30°E and SAT at every grid point in the North Atlantic region for (A) winter, (B) spring, (C) summer, and (D) autumn. Based on data for 1950–1989. Hatching and stippling for negative and positive correlations exceeding 95% confidence level.

winter averages between 90°W and 10°E were calculated for latitudes between 70°N and 30°N. These averages were smoothed by the cubic spline, and minimum and maximum values were taken as the latitudinal position of the Icelandic Low and Azores High respectively.

To assess the intensity and type of atmospheric circulation, indexes proposed by Katz (1960) and Pogosyan and Pavlovskaya (1977) were used.

Katz's westerlies index (I_w) characterizes mean intensity of the upper westerlies over the selected region and is calculated as

$$I_w = \left[\sum_{\phi = 30°N}^{65°N} \sum_{\lambda = 90°W}^{10°E} (H_\phi - H_{\phi + 5°}) \right] \Big/ n,$$

where H_ϕ and $H_{\phi + 5°}$ = 500 hPa heights at latitude ϕ and $\phi + 5°$ respectively, n = number of meridional gradients (only those meridional gradients were taken where $H_\phi > H_{\phi + 5°}$).

Pogosyan–Pavlovskaya's index (A) is based on computation of positive and negative anomalies of 500 hPa heights north and south of 50°N with subsequent subtraction of the second sum from the first. If negative anomalies prevail north of 50°N and positive ones south of this latitude ($A < 0$), the type of atmospheric circulation is assumed to be zonal. On the contrary, when $A > 0$, atmospheric circulation is considered as meridional.

Both indexes were computed for the region 70–30°N, 90°W–10°E.

To assess the degree of anomality of hydrometeorological fields the coefficients of anomality (K_a) proposed by Bagrov (Bagrov and Myakisheva, 1980) were computed as follows:

$$K_a = \sum_{i=1}^{n} \{[(x_i - \bar{x}_i) \cos \phi_i]/\sigma_i^2\} \Big/ \sum_{i=1}^{n} \cos \phi_i,$$

where x_i = value of SAT, SST, SLP, or H_{500} at grid point i, $\bar{x}_i$ = long-term mean, ϕ_i = latitude, σ_i = standard deviation, n = number of grid points.

Results

Atmospheric circulation

The decadal patterns of SLP and H_{500} anomalies for the last three decades are presented in Figure 3. During the 1980s negative anomalies of both SLP and H_{500} were predominant north of about 50°N, while south of this latitude they were mainly positive. This means that the 1980s as well as 1970s were characterized by strong westerlies both near the surface and in the upper layers, whereas anomalous easterly winds were a prominent feature of atmospheric circulation in the 1960s.

If we consider the value of the meridional pressure gradient, western flow was even more intensive during the 1980s than in the previous decade. In fact, as seen from Figure 4a, a marked tendency toward a strengthening of the upper westerlies was a characteristic feature of the 1980s. The highest values of Katz's index were observed in 1987 and 1989. At the same time, Pogosyan–Pavlovskaya's indices were negative in nine out of ten years (Fig. 4b), that is, indicative of a zonal type of atmospheric circulation during these years, as well as the whole 1980–1989 decade.

In contrast to the middle troposphere, there was no distinct sign of intensification of westerlies near the surface in the 1980s. On the contrary, as seen from Figure 5a, the atmospheric pressure gradient, the Azores minus Iceland, and, hence, the westerlies were stronger in the first half of the 1980s, becoming weaker in the second half of this decade (except for 1989).

In spite of the similarity of the decadal H_{500} and SLP patterns in the 1970s and 1980s there was an important difference between them, bearing in mind environmental conditions in the ICES area. Though westerlies over the North Atlantic were on average more intensive during the 1980s, orientation of isolines exhibited more pronounced southerly flows over the Northeast Atlantic in the 1970s. Some evidence of a weakening of warm air advection into the Norwegian Sea region from the south is given in Figure 5b. As can be seen, the intensity of southerly winds was near normal in the 1980s while they were anomalously developed in the 1970s.

The central pressure in the Icelandic Low (Fig. 6a) shows no significant trend in the 1980s. During recent decades its variations have resembled damping sine curve: the record high pressure at the end of the 1960s was followed by a drop to a record low value in 1974, when it tended to fluctuate near the normal.

Changes in latitudinal positions of the Icelandic Low and Azores High (Fig. 6b, c) are related fairly closely. The correlation coefficient between unsmoothed data is 0.62. Latitudinal position of the Icelandic Low fluctuates in a quasi-cyclic manner: in the first half of each decade it shifts northward and toward the end of the decade it tends to occupy a more southern position. As for the Azores High, during the 1980s it was situated in the northernmost position compared with the previous three decades.

Air and sea surface temperatures

In contrast to changes in global surface air temperature, the 1980–1989 decade was the coldest in the North Atlantic region. The most pronounced tendency toward cooling was observed in the northern North Atlantic (Fig. 7). Decrease of SAT started the 1950s and temperature reached the record low value in 1986 there.

When considering the spatical distribution of SAT anomalies during the last three decades (Fig. 8), one can see that their patterns were very similar in the 1970s and 1980s, in contrast to that observed in the 1960s. Negative

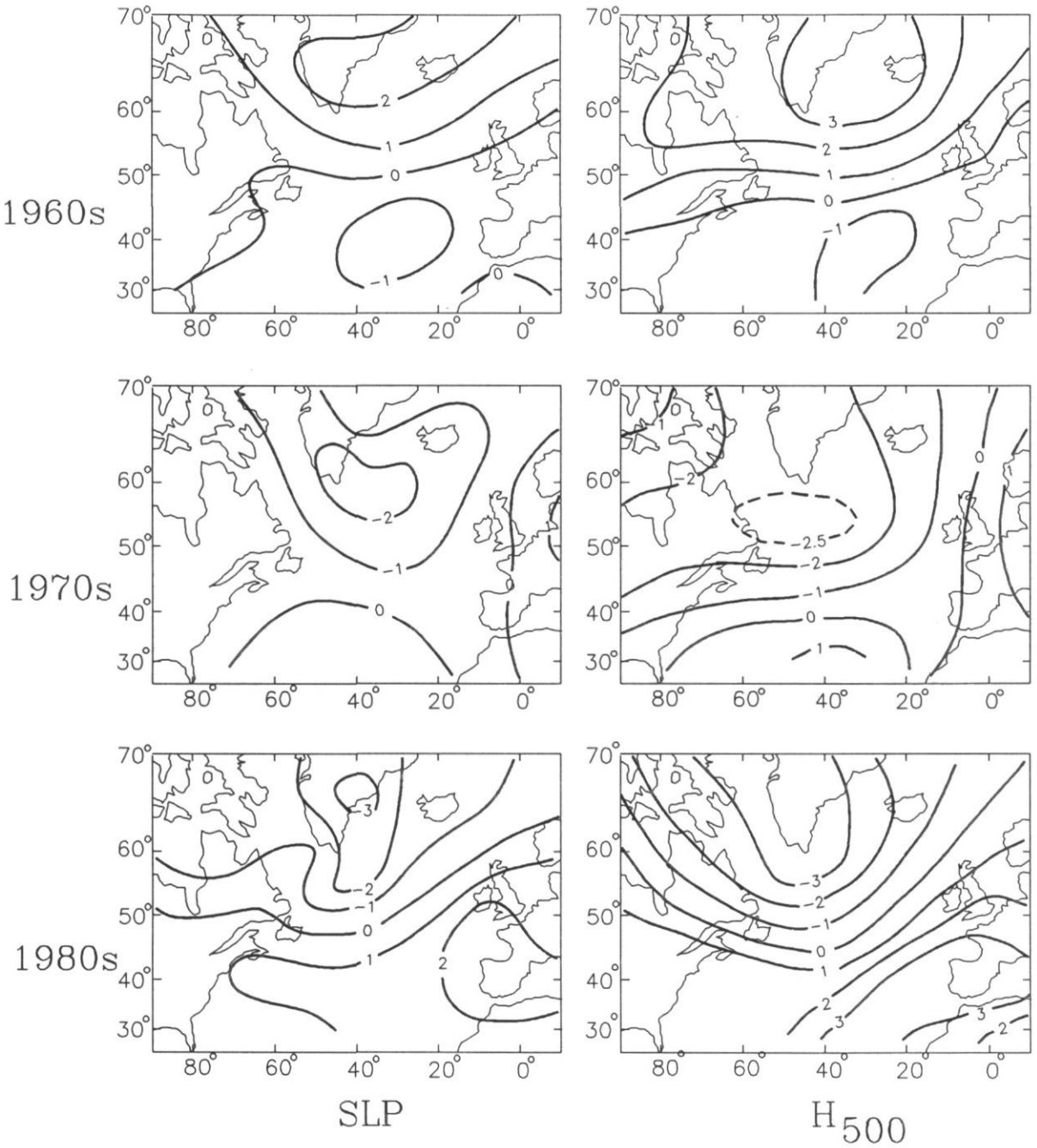

Figure 3. Mean winter sea-level pressure (SLP) and 500 hPa heights (H_{500}) anomalies for decades 1960–1969, 1970–1979, and 1980–1989. Anomalies of SLP are in hectopascales and those of H_{500} in geopotential meters. Departures from long-term mean for 1951–1980.

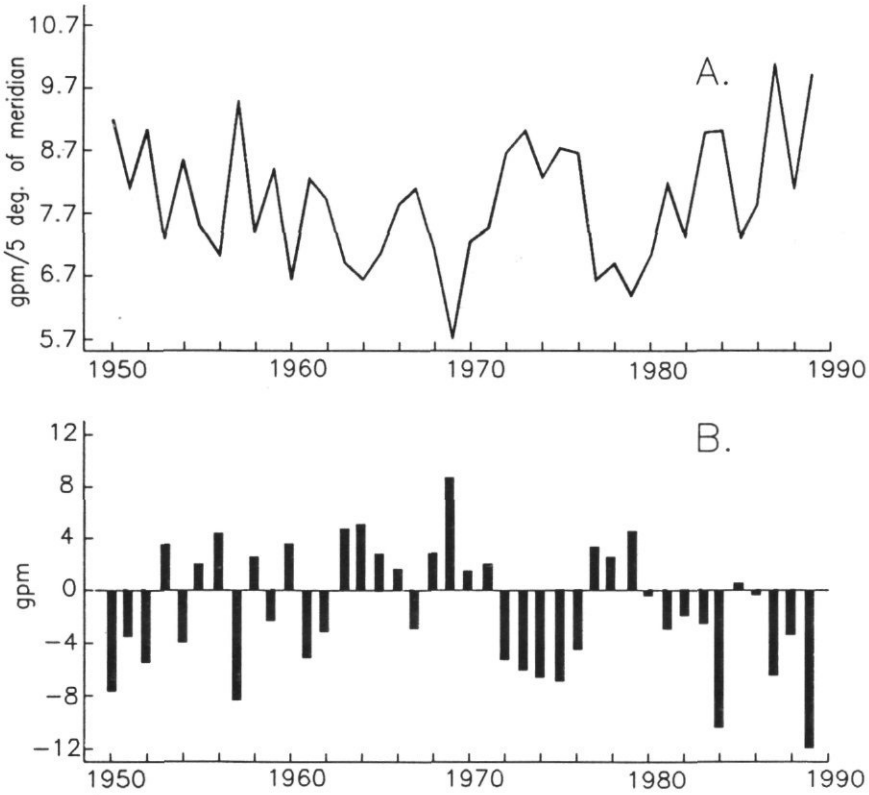

Figure 4. Changes in (A) Katz's index and (B) Pogosyan–Pavlovskaya's index for winter season in the North Atlantic region.

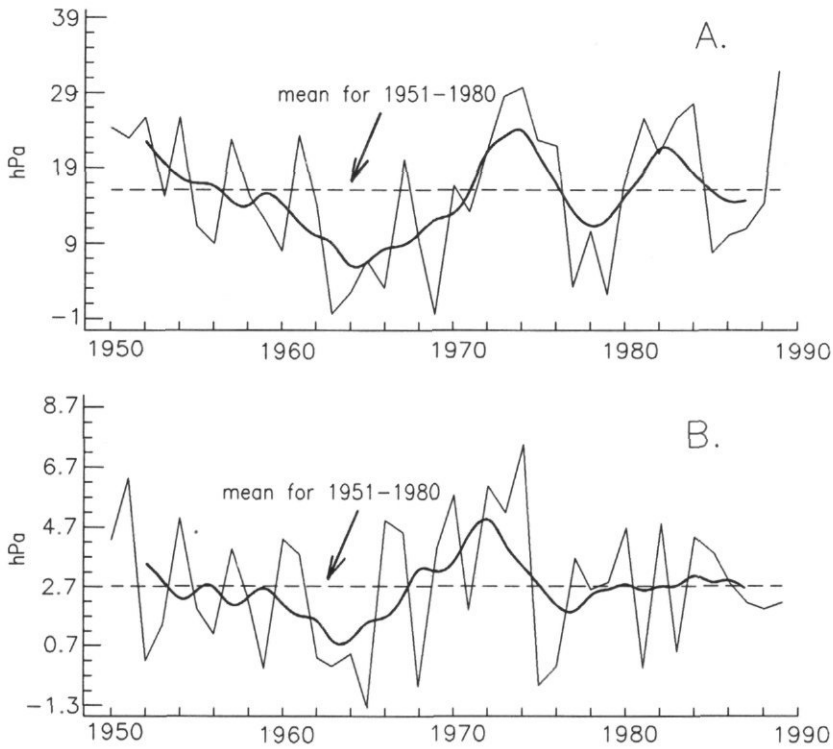

Figure 5. Sea-level pressure (SLP) differences between (A) Azores (35°N 30°W) and Iceland (65°N 20°W) and (B) grid points 60°N 10°E and 60°N 0°E for winter season. Solid curves represent five-year running means.

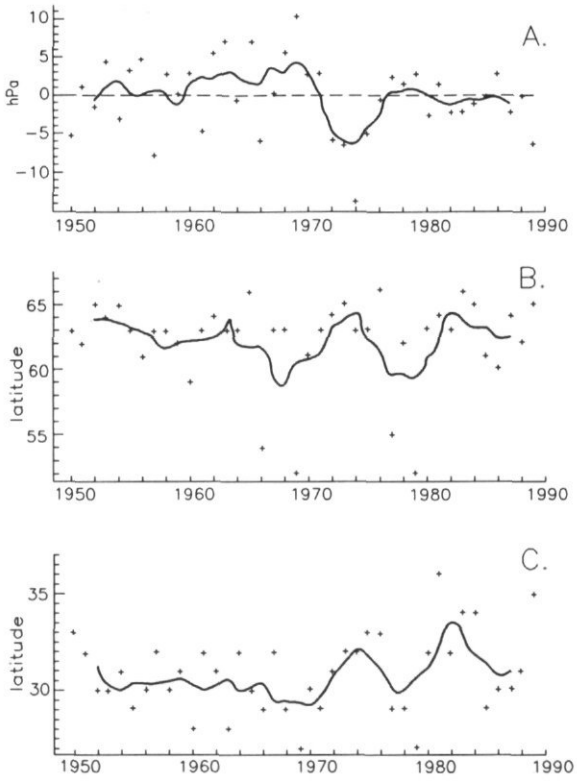

Figure 6. Changes in (A) central pressure of Icelandic Low (departures from long-term mean in hectopascales) and latitudinal positions of (B) Icelandic Low, and (C) Azores High for winter season. Solid curves are five-year running means. Crosses represent winter means.

air temperature anomalies occupied almost the whole North Atlantic with the lowest values in the western ocean in both the 1970s and the 1980s.

One of the most striking features during the last decade was a very large positive air temperature anomaly over Greenland. Its value reached 6.5°C, even for decadal averages. According to data available in the archives of the USSR Hydrometeorological Center, a sharp, stepwise increase in air temperature occurred from 1969 to 1970, when mean winter temperature rose to 9.7°C (at grid point 70°N 40°W), and then continued to fluctuate around a new, higher mean level. The abrupt change in air temperature coincided with a shift in character of atmospheric circulation, as shown above. A similar sharp increase in air temperature in western Greenland was observed during the 1920s (Rogers, 1985).

Mean decadal SST patterns are also shown in Figure 8. Like SAT, SST patterns were similar to each other during the 1970s and 1980s, with a predominance of negative anomalies in the northern North Atlantic, and differed from that in the 1960s. However, if during the 1970s negative anomalies were observed in 73% of grid points, in the 1980s they occupied only 37% of the area covered by data, though their absolute values were higher. The most notable warming during the 1980s occurred in the Gulf Stream area, where mean decadal sea surface temperature anomalies reached 0.7°C.

Discussion

Several periods with different climatic regimes can be revealed in fluctuations of hydrometeorological variables in the North Atlantic during recent decades. The time scale of such regimes is assessed to be about 5–15 years (Namias, 1982). Changes in annual sea temperature at the Kola section, which is indicative of thermal conditions in the whole Northeast Atlantic, are a good example of the transition from one regime to another (Fig. 9). As seen, there have been three well-pronounced thermal regimes over the last 30 years. The cold regime existing in the 1960s was replaced by the warm one of the first half of the 1970s (Dickson *et al.*, 1975). In 1977 another shift occurred, which, as in previous cases, was accompanied by radical changes in the whole atmosphere–ocean system of the North Atlantic. A new climatic regime existing in 1977–1982 was similar to that during the 1960s (Dooley *et al.*, 1984; Rodionov and Krovnin, 1990). The particular importance of events occurring in the climate system around 1977 comes from the fact that they resulted in the end of a tendency toward a decline of the Caspian Sea level observed over many years and the start of an unusually rapid rise (Rodionov, 1989). The Caspian Sea level may be considered as a natural low-pass filter smoothing short-period climate fluctuations, so reversal in its tendency reflects the drastic changes in the whole climate system.

It should be noted that similar regimes are distinguished in fluctuations of hydrometeorological characteristics in widely separated regions of the Northern Hemisphere. So, Namias (1982) found the same regimes in variations of SAT in the southeastern part of the United States. They were also revealed in the Bering Sea (Rodionov and Krovnin, 1991).

An important question arises concerning whether a new climatic regime has been established since about 1983 similar to that of the first half of the 1970s. The positive answer to this question sounds credible if one takes into account that 1982 was a year of a very strong El-Nino event which many scientists consider a signal of drastic changes in the global climate system. However, there was no essential change in the variations of many of those hydrometeorological characteristics in which the above regimes were displayed clearly in previous years. For example, the thermal regime established in the Bering Sea in 1977 continued to be warm during the 1980s, and the Caspian Sea level continued to rise though at a somewhat slower rate.

As indicated above, mean decadal thermal and circulation characteristics in the North Atlantic region were

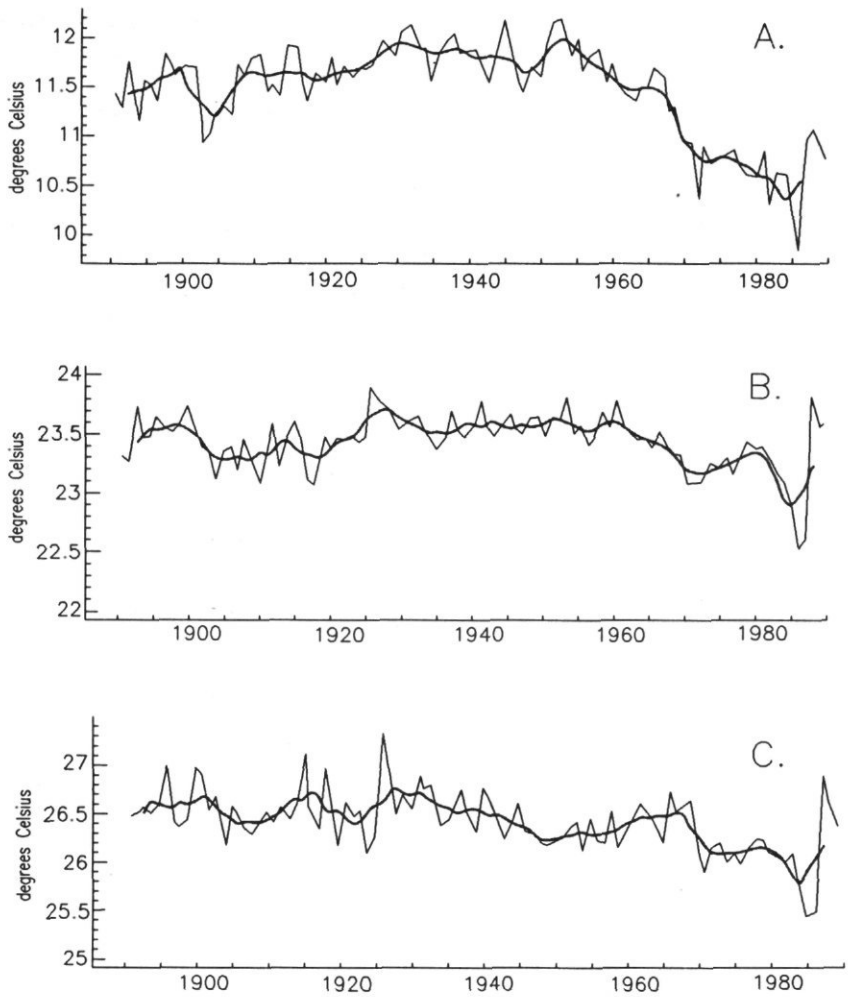

Figure 7. Annual surface air temperature over the North Atlantic averaged for latitudinal zones: (A) 37.5–67.5°N, (B) 12.5–37.5°N, and (C) 2.5–12.5°N. Adapted from Gruza *et al.* (1990).

similar during the 1970s and 1980s, but there were some essential differences. In contrast to the first half of the 1970s, the predominance of zonal atmospheric circulation did not result in a marked increase of SST in the Northeast Atlantic in the 1980s, though 1983 and 1989 were very warm years there (Fig. 9).

The important feature of the 1980s is an increase in degree of anomaly of many hydrometeorological processes. Table 1 shows that Bagrov's coefficients of anomaly for SAT and especially SLP and H_{500} were much higher in the 1980s compared with previous decades.

Some evidence of an increase in interannual variability during the last forty years is produced by variations in index of the NAO (Fig. 10). Variance of fluctuations in the NAO index was five times higher during the 1970–1989 period than during the 1950–1969 period. The index ranged from −3.90 (1986) to 3.93 (1989) in the second half of the 1980s, i.e. from a record low to a record high value.

Winter 1986 was very mild over and west of Greenland. Mean winter air temperature anomalies exceeded standard deviation by a factor of 2. At the same time, large negative anomalies were observed over the rest of the North Atlantic. This winter was also very severe in northern Europe.

Advection of warm air into the Greenland area was associated with an establishment of a pronounced high-

Table 1. Bagrov's coefficients of anomaly for mean decadal anomaly fields of sea-level pressure (SLP), 500 hPa heights (H_{500}), surface air temperature (SAT) and sea-surface temperature (SST).

Decades	SLP	H_{500}	SAT	SST
1950s	0.96	1.52	0.77	—
1960s	0.96	1.55	0.60	0.26
1970s	0.81	1.50	0.99	0.18
1980s	1.48	2.35	1.17	0.24

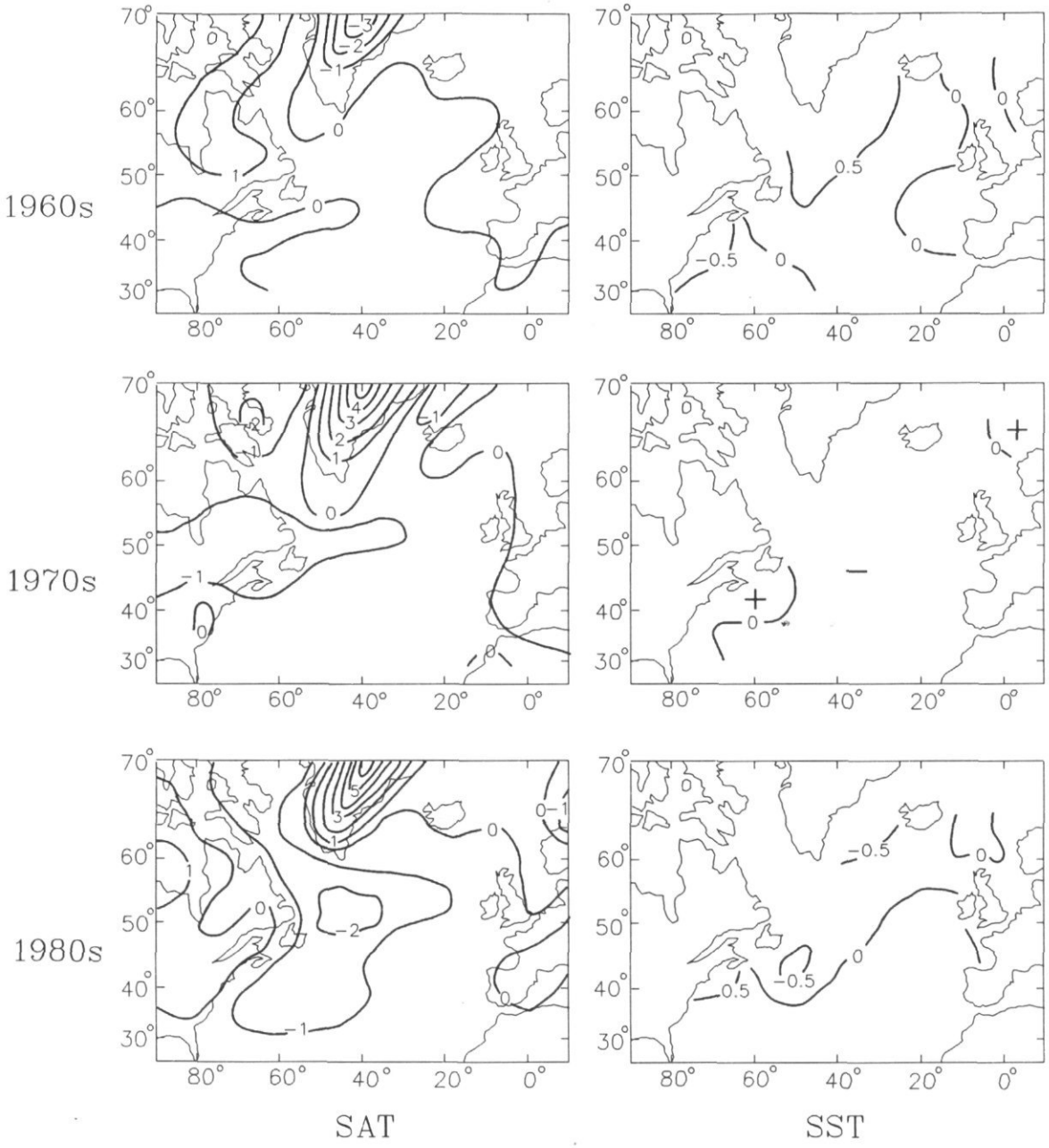

Figure 8. Mean winter air and sea surface temperature anomalies (°C) for decades 1960–1969, 1970–1979, and 1980–1989. SAT anomalies are deparatures from long-term mean for 1951–1980 and SST anomalies are departures from long-term mean for 1957–1989.

pressure cell east of Greenland, while large negative pressure anomalies were observed over the northeastern part of Canada. Such SLP anomaly patterns resulted in strong southerlies over the Greenland–Labrador area. At the same time, northerly outbreaks of cold Arctic air swept Western Europe.

A contrasting situation was observed in winter 1989. Intensive advection of warm Atlantic air led to a very mild winter in northern Europe. In Norway, anomalies of SAT exceeded the standard deviation by a factor of 3. Strengthening of westerlies over the North Atlantic was accompanied by northerly outbreaks west of Greenland.

It should be noted that during the 1980s an increase in variability was observed not only at the interannual time scale but at the intraseasonal time scale too. A good example of the latter is winter 1988, the mean thermal

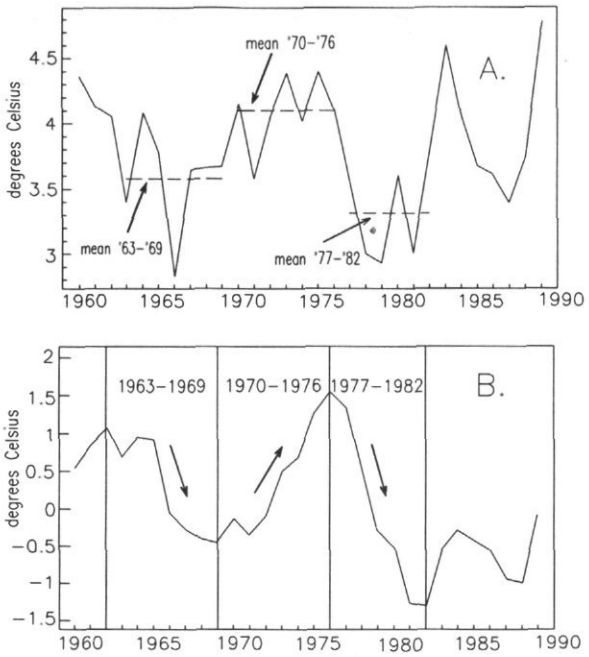

Figure 9. Thermal regimes in the Northeast Atlantic as shown by changes in (A) annual sea temperature and (B) cumulative sum of its anomalies in the 0–200 m layer at Kola section.

and circulation characteristics were near normal. However, as shown by Rodionov and Pogodin (1990), the range of variability in atmospheric circulation during this winter was exceptionally wide: from anomalously developed cyclonic activity at the end of December 1987 to the beginning of January 1988 to the establishment of a strong blocking anticyclone in the second half of February 1988. SLP patterns for two five-day periods, 27–31 December 1987 and 25–29 February 1988, are presented in Figure 11. In the first case negative anomalies in the centre of the cyclone exceeded 25 hPa (relative to the 1951–1970 reference period). Such a low pressure during this five-day period was not observed over the period 1951–1970, which is characterized by abnormally

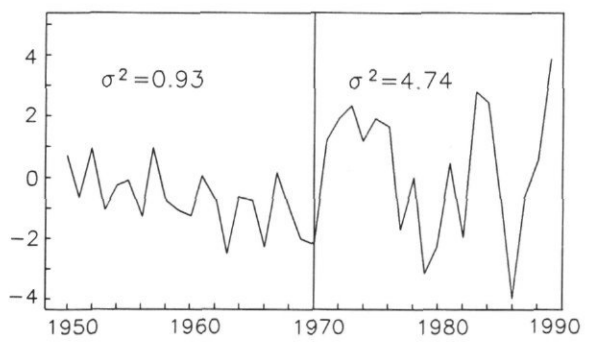

Figure 10. Index of the North Atlantic Oscillation taken as difference between normalized mean winter surface air temperature anomalies at grid points 60°N 10°E and 70°N 50°W.

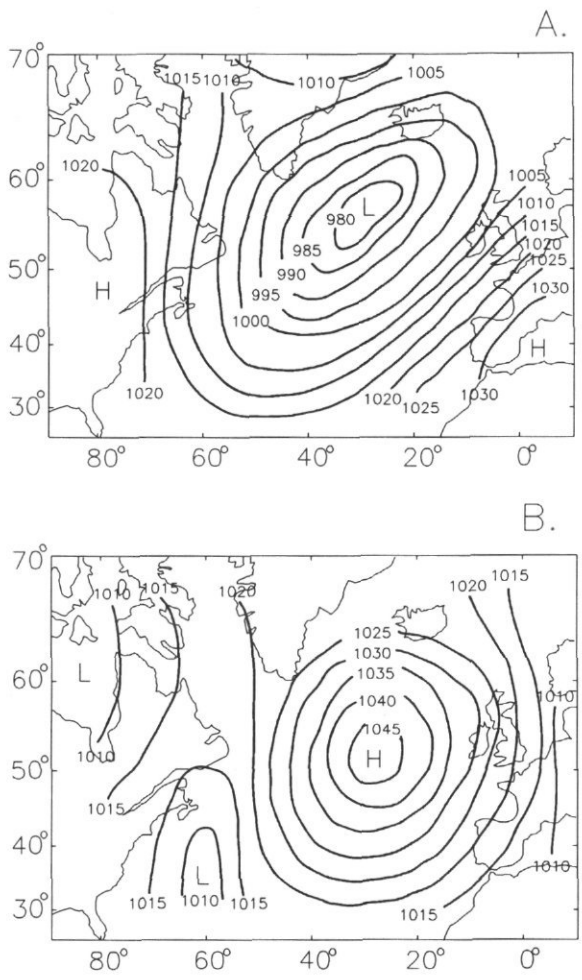

Figure 11. Mean sea-level pressure (hPa) in the North Atlantic for two five-day periods: (A) 27–31 December 1987 and (B) 25–29 February 1988.

high pressure north of 50°N and frequent occurrence of blocking anticyclones, nor during the period 1919–1938, when the Icelandic Low was very deep and a zonal circulation pattern prevailed (Perry and Fearnside, 1980). The situation established by the end of February 1988 (Fig. 11b) was the complete opposite to that observed two months previously. Central pressure in an anticyclone was exceptionally high (1045 hPa), i.e. 35 hPa higher than long-term mean. Such a strong anticyclone over the central North Atlantic was not observed in any five-day period (not just in winter but even in summer months) over the periods 1890–1909, 1919–1938, and 1951–1970, data for which were presented by Perry and Fearnside (1980).

It may be supposed that an increase in climatic variability is not a regional feature peculiar to the North Atlantic, but perhaps also occurs in other regions of the Northern Hemisphere and as a result is reflected in an increase in interannual variability of mean air tempera-

ture for the whole Northern Hemisphere. In fact, as seen from Figure 1, ranges between both upper and lower quartiles and maximum and minimum values of SAT in the Northern Hemisphere were the highest in the 1980–1989 decade.

Summary

Against a background of global warming the decade 1980–1989 was the coldest in the North Atlantic since the end of the last century. A decrease in air temperature, most pronounced in the northern North Atlantic, started as early as the 1950s and had reached its peak by the middle of the 1980s. In recent years temperature has tended to return to normal.

The 1980s are characterized by the development of strong westerlies over the North Atlantic that as a rule are accompanied by increased head advection into the Northeast Atlantic and the corresponding rise in air and sea surface temperatures. However, in contrast to the 1970s, the clear tendency toward warming was not observed in this region during the 1980s.

The most prominent feature of the 1980s is the increase in variability of hydrometeorological processes at both intra- and interannual time scales. Both the record low and high values of the NAO index were registered in the 1980s.

According to Barnett (1985), NAO appears capable of exhibiting two different temporal behaviours, "multiple-equilibria" and "continuous". In one form, the NAO pattern assumes a character of "climatic state", which, once established, maintains its characteristics for years. In the second form, it appears as a continuous function of time. During the last 30 years it is possible to distinguish three climatic states of the NAO which resulted in three well-pronounced climatic regimes in the Northeast Atlantic: 1963–1969, 1970–1976, and 1977–1982. During the rest of the 1980s the NAO tend to exhibit a continuous form of behaviour with a very large range of characteristics.

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Long-term variability of the in–out flow to the North Sea

G. K. Furnes

Furnes, G. K. 1992. Long-term variability of the in–out flow to the North Sea. – ICES mar. Sci. Symp., 195: 103–109.

This paper deals with the wind-induced variability of the transport fluxes in the Norwegian Trench. Calculations are made of monthly mean flux values from January 1881 to December 1989. The resulting time series is subsequently examined with respect to events with a duration of several years. It is shown that the fluxes alternate between periods of high (enhanced outflow) and low (weakened outflow) values. It is suggested that this variation is reflected in the hydrographic relations of the North Sea.

G. K. Furnes: Norsk Hydro Research Centre, PO Box 4313, 5028 Bergen, Norway.

Introduction

The oceanographic conditions in the North Sea are subjected to variations on many time scales. Some of the variations are a result of regular forcing mechanisms, others may be responses to stochastic processes. The most relevant factors causing the variations are likely to be: variations in wind-forcing; variations in runoff and precipitation/evaporation; variations in the heat exchange; variations in the ocean boundary conditions.

This paper deals with the effects of wind-forcing, namely the computation of long-term wind-induced transport fluxes in the Norwegian Trench. In earlier papers (Riepma, 1978; Dooley and Furnes, 1981) it has been demonstrated that the pronounced changes that appear in the Norwegian Trench outflow are present in the Atlantic inflow on the North Sea plateau as well. Increasing outflow in the Trench leads simultaneously to increasing Atlantic inflow on the plateau, and vice versa (for details on the dynamical mechanisms, see Furnes (1980) and Davies and Heaps (1980)). The Atlantic inflow is of great importance for the heat and salt budget of the North Sea. Long-term variations that may occur in the in–out flow to the North Sea are therefore expected to be reflected in various environmental parameters as well.

Numerical simulation models have successfully been used to reproduce observed fluxes in the Norwegian Trench (Backhaus, 1990; Furnes, 1991). However, the observation series are few and of short duration, typically a couple of months. In lieu of longer series, it is believed that transport fluxes can be achieved from numerical simulation models with satisfactory accuracy, as in Backhaus (1990), where transport fluxes through a section in the Norwegian Trench are computed for more than 10 years.

In the present paper, the fluxes are computed as monthly mean values, from January 1881 to December 1989. For such a long time period, the numerical simulation models which are based on the hydrodynamical equations would involve large computing resources. A simplified description using a regression approximation has therefore been applied in the following study.

Data and methods

The formula used to calculate the long-term fluxes was determined from a multiple regression analysis in which the fluxes are expressed as a linear combination of atmospheric pressure and wind-stress components in the east–west and north–south direction. In order to determine the regression coefficients, the observed fluxes through the Utsira section (Furnes and Sælen, 1977) were used together with wind data from the central northern North Sea. The flux data cover a period of nearly two months in March and April 1976. Having established this formula it is subsequently assumed that it can be used to calculate transport fluxes through the same section, provided that wind data are available from the same location.

Time series of wind and pressure have been composed from two databases. Before 1955, the data source is the World Meteorological Organization (WMO) and from January 1955 it is from the Norwegian Meteorological Institute (DNMI). The two databases overlap from 1955 to 1982. By comparing the time series in the overlapping period, a small correction of the WMO data has been

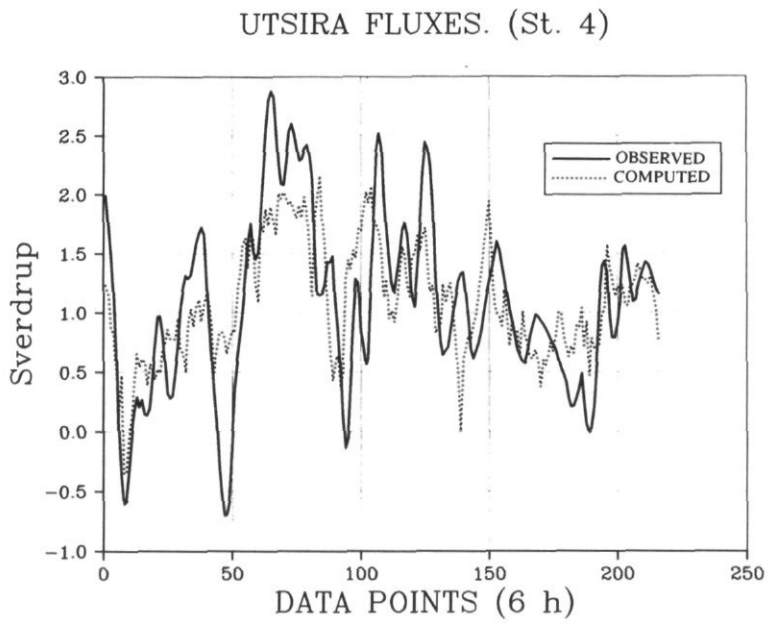

Figure 1. Observed transports in the Utsira section together with computed values from a regression model.

introduced to establish a long and homogeneous time series (Furnes and Reistad, 1991). The location of the gridpoint in the database represents a position in the central northern North Sea. In the calculation of wind stresses in the regression formula, a quadratic law was applied following the formulation of Large and Pond (1981).

An impression of how the regression formula fits the observed Utsira fluxes may be gained from Figure 1. The data are plotted with 6 h resolution and positive flux values indicate flow towards the north. It is seen from the figure that deviations from the observations are most pronounced on the shortest time scales, while the longer scales seem to be better rep-

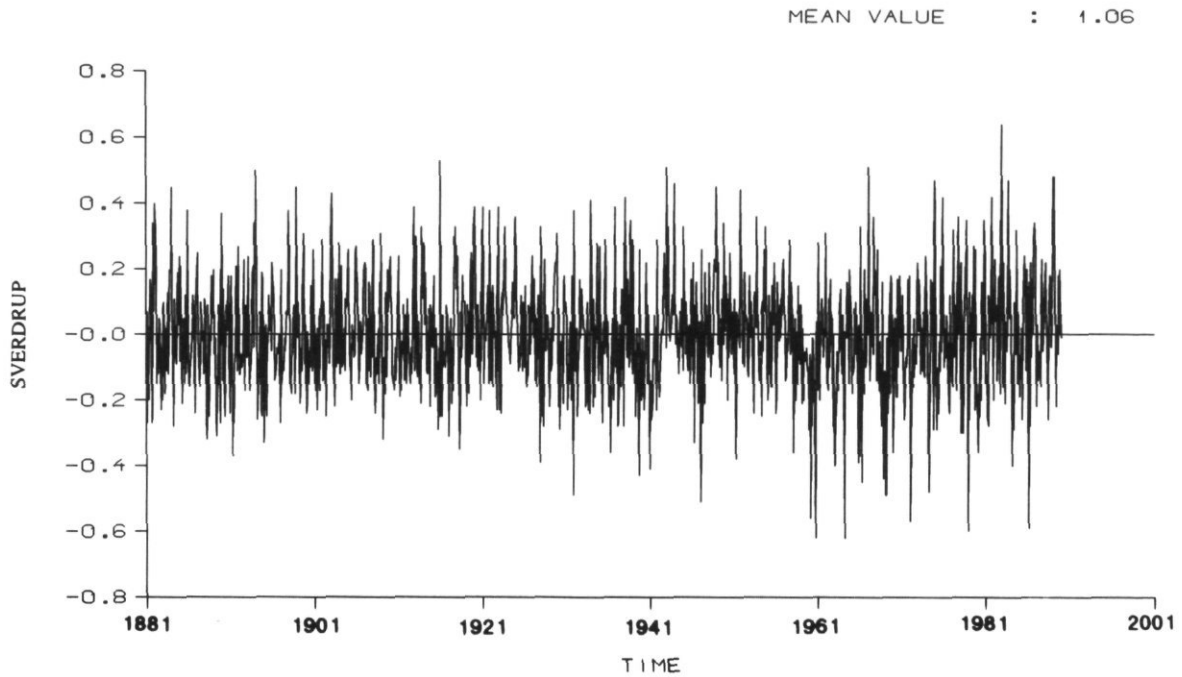

Figure 2. Time series of computed monthly mean transports.

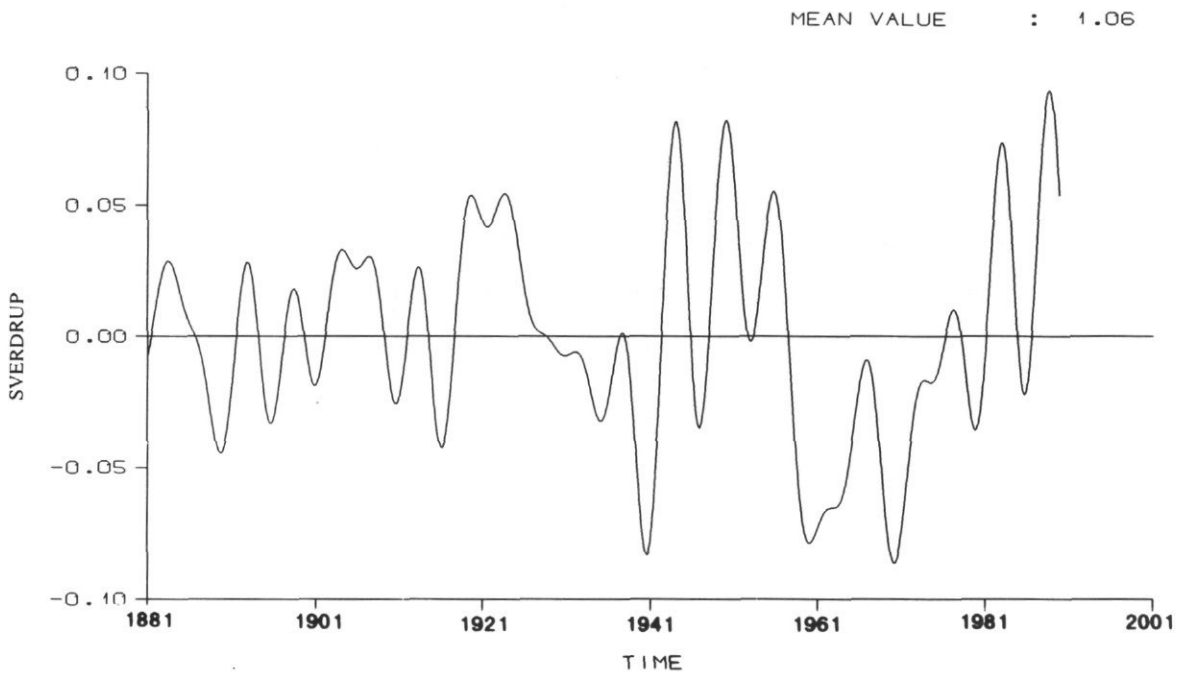

Figure 3. Low-passed filtered version of Figure 2. Cut-off 0.2 year^{-1}.

resented. The computed correlation coefficient was about 0.7.

Assuming the regression formula can be applied to the monthly mean values of wind forcing, time series of transport fluxes representing the Utsira section are computed from January 1881 to December 1989 and plotted in Figure 2. It is evident from this figure that the fluxes have an extensive variability around the mean value of 1.06 Sv. Positive deviation from the mean value represents enhanced outflow (north), and negative weakened. Most of the variability appearing in the figure is of seasonal origin.

In the following, long-term variations are emphasized by suppressing events with time scales shorter than five years. This is achieved by running the monthly mean flux data through a low-pass filter with cutoff frequency of 0.2 year^{-1}. The new time series is plotted in Figure 3.

Results and discussion

Referring to Figure 3, it is interesting to note some of the pronounced long-term events appearing in the transport fluxes in the last 109 years. The first part of the time series up to the beginning of the 1920s seems to indicate a tendency towards increasing outflow and hence increasing inflow of Atlantic water. From the early 1920s and towards 1940 there is a gradual reduction again, leading to one of the lowest fluxes of the whole series. Between the first part of the 1940s and the mid-1950s, there is a dominance of enhanced outflow. This event is followed by the most peculiar period in the time series. From the end of the 1950s to the end of the 1970s, the most prominent negative anomaly is found in the computed trench fluxes. The lowest values seem to take place about 1969/70. During the 1980s there is again a gradual enhancement of the outflow, which in magnitude is similar to that of the 1940s. The highest computed flux values are found at the end of the period.

The changes of strength of the Atlantic inflow for the years 1970–1989 have been estimated by van de Kamp (1990). The calculations are based on charts of the February distribution of bottom salinity. Verification of the computed long-term transport fluxes presented in this paper is difficult to achieve. With access to regular long-term hydrographic observations, it would be interesting to examine the salinity data for possible coinciding anomalies. However, the most regular salinity observation which is available is data from Flødevigen at the southern coast of Norway. Since 1919 salinity has been measured daily at 1 m depth. Figure 4 presents the five-year running means of the yearly mean salinity, as presented by Ljøen and Sætre (1987). In Figure 5 the low-passed (cutoff 0.2 year^{-1}) transport fluxes through the Utsira section for the same period are shown. Note the different ways in which the two time series are filtered. By comparing Figures 4 and 5 it is seen that increasing or decreasing fluxes coincide roughly with increasing or decreasing salinities. The causal connection, however, is not clear as local surface processes are also expected to influence the salinity data.

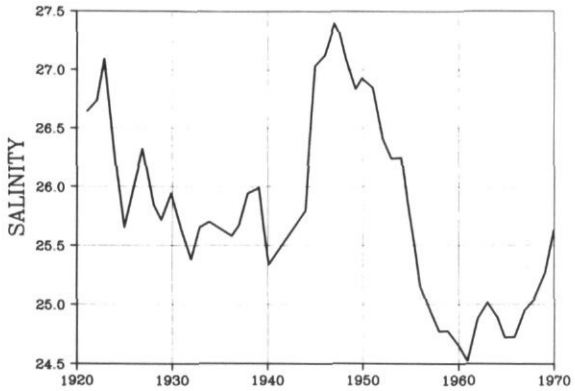

Figure 4. Five-year running mean of yearly salinity means at Flødevigen.



Figure 5. Computed low-passed transports at Utsira, extracted from Figure 3.

An important question is how the fluxes evolve at different seasons. In Figures 6 and 7 average fluxes for January–June and for July–December respectively, are shown. The averages are low-pass filtered with a cutoff of 0.1 year^{-1}. The figures show that the time evolution of the fluxes differs considerably. Some events, however, seem to affect both half-years, as e.g. the positive anomaly at the beginning of the 1920s and the minimum in the late 1950s. One of the striking differences between the figures is the different development after 1960. The negative anomaly seems to persist much longer for the January–June averages than for the second half-year, which exhibits a steady increase up to about 1982.

To give a more detailed presentation of the fluxes in recent years, quarterly averages have been plotted in Figures 8–11 from 1967–1989. As seen in Figures 10 and 11 the steady increase in fluxes during the second half-year is evident in both the July–September and the October–December averages. In particular the July–September values seem to have a steady trend towards the end of the 1980s.

Comparing Figures 8–11 it is evident that the strongest mean flow takes place during October–December. On average it exceeds 1.1 Sv in this period. However, the highest single quarterly mean value is found in January–March 1989. In contrast, the lowest value also appears in the January–March average, in which two events have transport values below 0.7 Sv, namely in 1969 and 1972. Individual events or events with shorter duration frequently have much lower transports, as seen in Figure 2,

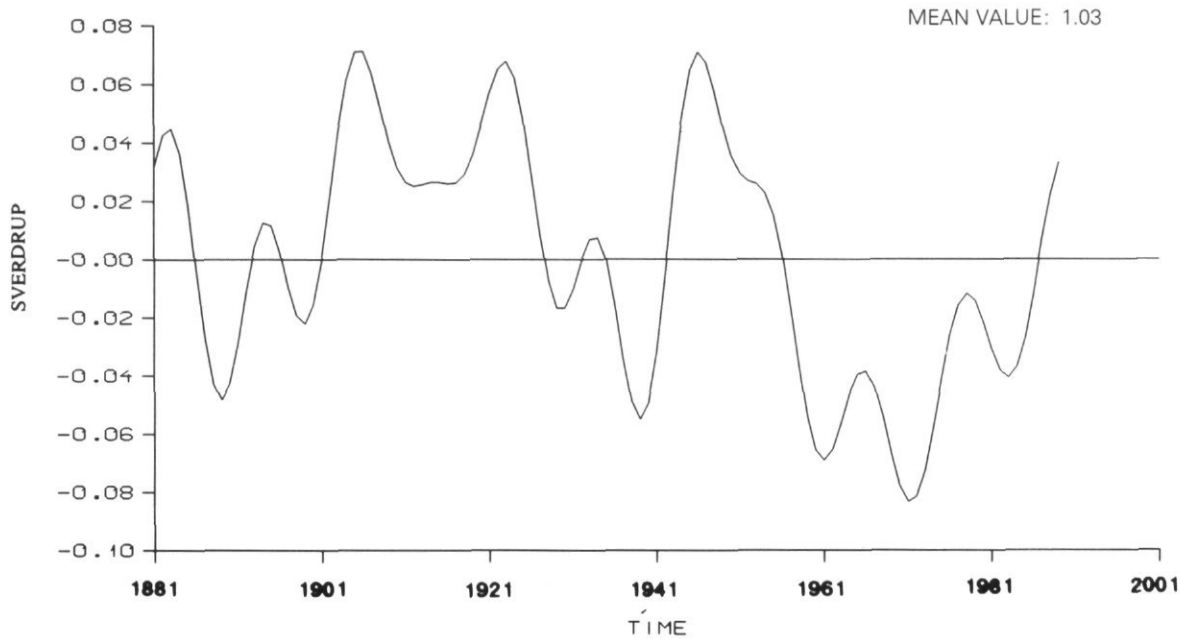

Figure 6. Time series of low-passed mean transports for January–June. Cut-off 0.1 year^{-1}.

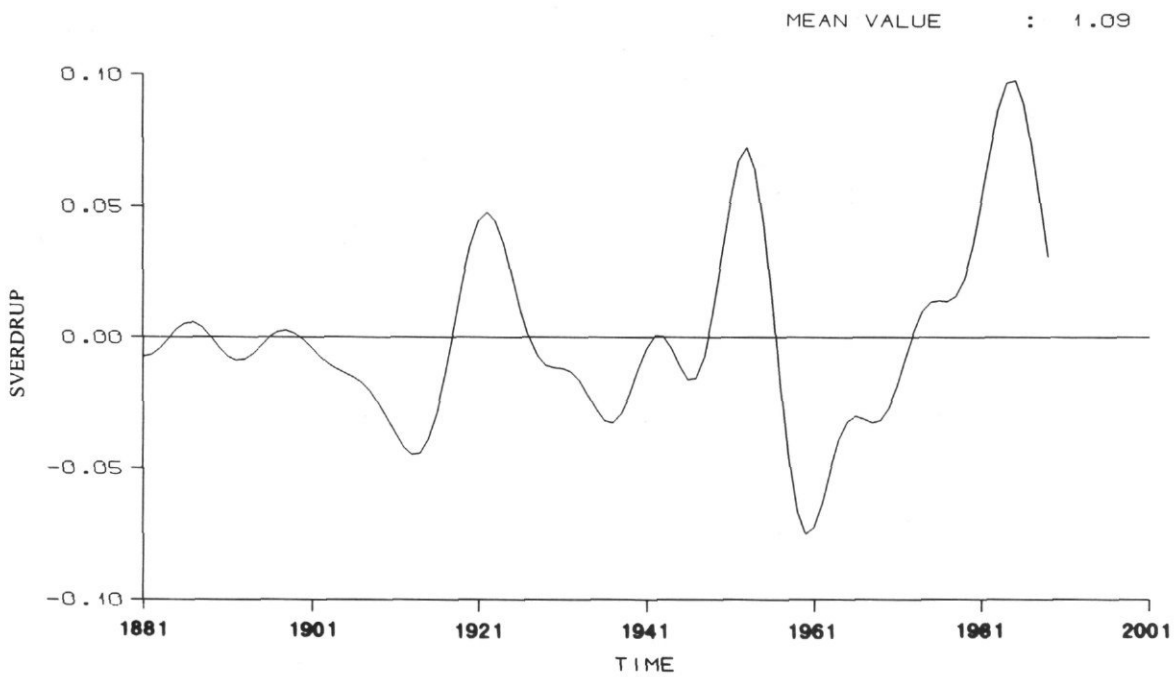

Figure 7. Time series of low-passed mean transports for July–December. Cut-off 0.1 year^{-1}.

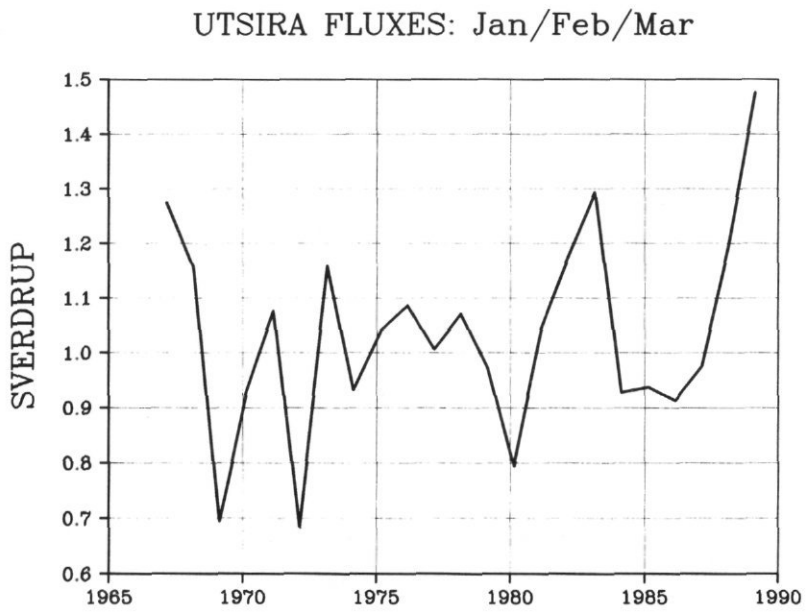

Figure 8. Quarterly mean values of transports from 1967 to 1989 based on January to March averages.

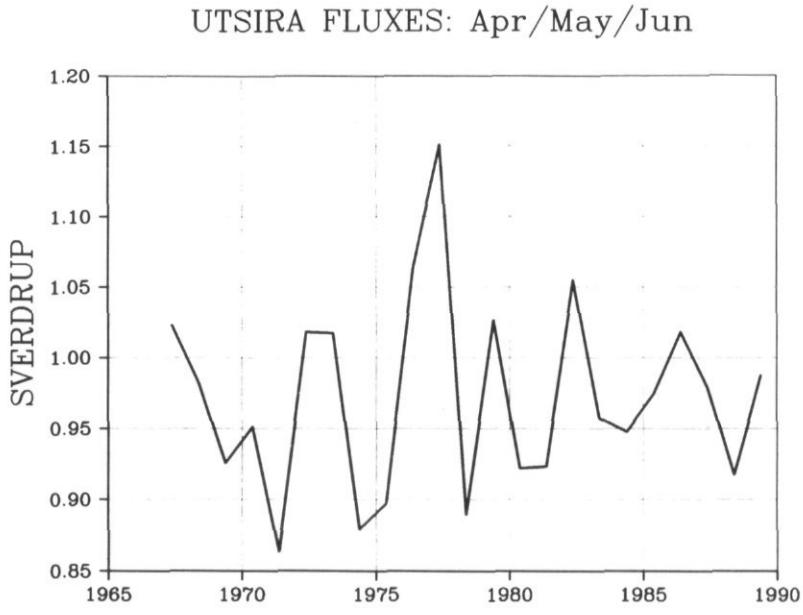

Figure 9. As for Figure 8 for April to June.

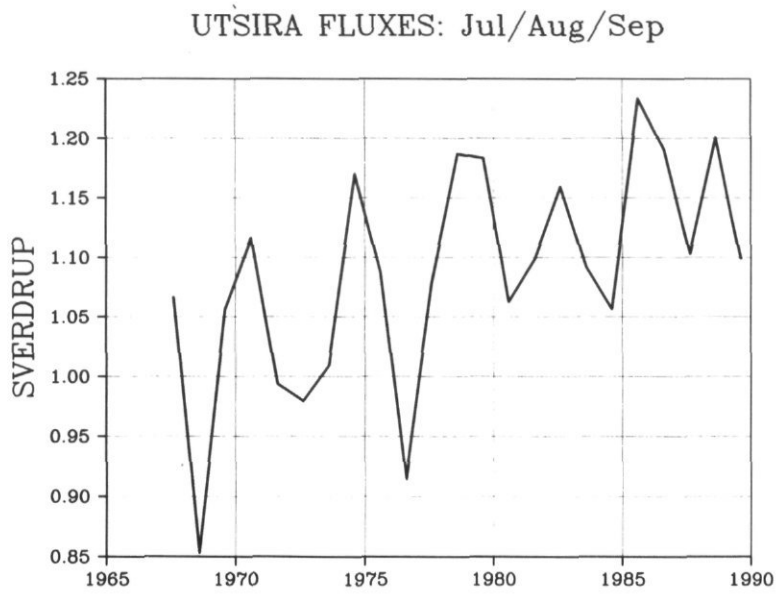

Figure 10. As for Figure 8 for July to September.

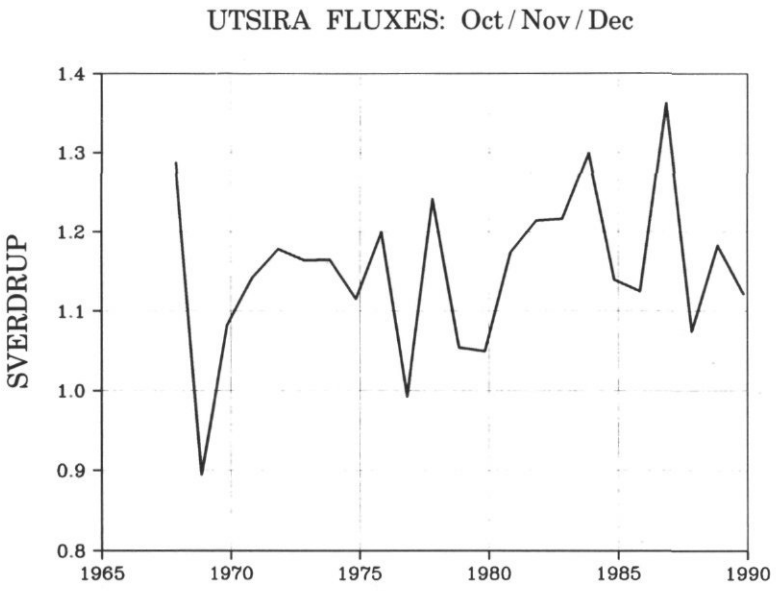

Figure 11. As for Figure 8 for October to December.

especially after about 1960. Regarding the averages for April–June, the most pronounced flux value is the maximum that appears in 1977.

As mentioned earlier, it is well established through theoretical and experimental studies that there is a strong coupling between the large-scale wind field over the North Sea and the transport fluxes in the Norwegian Trench. In fact it has been demonstrated that the wind forcing is the dominating signal in the residual current records. The formulation of a regression model, from which the above transports are deduced, includes the wind forcing terms which according to the mentioned literature have been regarded as important for the flow variability. It is therefore suggested that the present model formulation can resolve the major events that have taken place in the considered period.

However, it is also clear that other mechanisms influence the strength of the fluxes in the Norwegian Trench. The most important are the baroclinic processes, which are not taken into account in the calculations that have been presented here.

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Temporal and spatial trends of organic micropollutants in the sea water of the Baltic Sea, the North Sea, and the Northeast Atlantic

Horst Gaul

Gaul, H. 1992. Temporal and spatial trends of organic micropollutants in the sea water of the Baltic Sea, the North Sea, and the Northeast Atlantic. – ICES mar. Sci. Symp., 195: 110–126.

Horst Gaul: *Federal Maritime and Hydrographic Agency, PO Box 30 12 20, D-2000 Hamburg 36, Germany.*

Introduction

Oceanographic data have a variability in both space and time. Periodic time scales range from a fraction of a day to several thousand years (Table 1). All periodic events interfere and overlap with additional aperiodic single incidents, thereby resulting in a complicated pattern.

The organic micropollutants discussed here are all man-made products. They do not have natural background values that could camouflage low-level gradients and are thus suitable tracers for hydrographic studies. Hexachlorocyclohexane (HCH)-isomers are found throughout the ICES area because they survive long periods of transportation and are also soluble enough to escape sedimentation.

Technical HCH (with 60% α-HCH, 15% γ-HCH (lindane) and 10% β-HCH) came into general use in 1950. Since 1980, however, use of γ-HCH in Europe has been allowed only as an insecticide.

The use of hexachlorobenzene (HCB) as a fungicide and seed-dressing was banned in the mid-1970s. Although it is no longer applied, it reaches the environment as a by-product of incineration or chlorination. HCB is not found offshore because it is rapidly trapped in the sediment.

The method applied (Gaul and Ziebarth, 1983) consists of the following items.

- sampling with an all glass and stainless steel sampler
- hexane extraction of unfiltered water within the sampler
- clean-up of the extract with sulphuric acid
- gas chromatography on capillary columns with electron capture detector (ECD)

The quantification limits of the method depend on compound specific procedural blanks and detector responses and range from 0.02 to 0.05 ng/l. At levels well above the quantification limit (tenfold or more), the relative standard deviation is below 10%.

Table 1. Natural periods.

Cycle	Period	Example
Tide	~12.4–24.8 h	Salinity in estuaries
Light	24 h	Nutrients
Phase inequality	14.8 days	Neap and spring tide
Seasons	1 year	Nutrients
?	3 years	Abundance of *Noctiluca scintillans* (Uhlig and Sahling, 1990)
Sunspot activity	~11 years	T/S data (?)
Half cycle of Earth's precission	~11 000 years	Ice-rafted debris in N Atlantic sediments (Heinrich, 1988)
Tilt of Earth's axis	~40 000 years	Climate
Eccentricity of Earth's orbit	~100 000 years	Climate

Temporal trend

There are no observations of variations in these pollutants on a daily cycle extending over the analytical margin of error of the method. In estuaries in the German Bight, however, we find the influence of the tide on the concentration of some contaminants, but it becomes less significant the nearer we approach the open sea (Fig. 1).

For the pesticide lindane, a seasonal dependence was expected. The maximum in July, however, is more likely

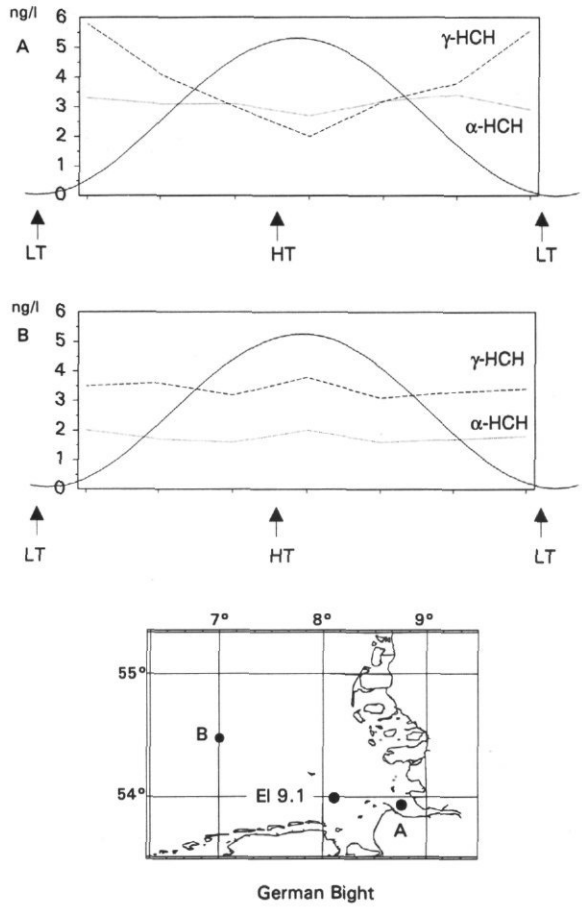


Figure 1. Correlation of concentration of HCH's with the tide: (A) in Elbe Estuary, (B) and the open sea.

the result of the low drainage volume of rivers in the summer than of lindane application in agriculture (Fig. 2).

German Bight

For long-term trend assessment, we compiled data at least once a year from 1980 onwards from a station near the Elbe light vessel in JMP area 13. The data can be compared as personnel and methods have not changed in that time (Fig. 3). The station is a compromise between the highly variable concentrations in estuaries and the decreasing variability and concentrations in offshore waters.

The station is regularly *not* within the plume of the Elbe; the salinity varies mostly between 30 and 32.5 and seldom drops below 28 PSU.

The analytical data were not fully supplemented with T/S data, so a normalization for salinity was not possible. Statistical methods were not applied because the amount of data is too small. The time series at the station shows a decrease for α-HCH and HCB, but not a clear trend for lindane. The very low values in 2/81 and 3/81 are not felt to be representative of normal flushing conditions in the North Sea.

The decrease in α-HCH is the result of the ban on the use of technical HCH as an insecticide, of which approximately 60% was α-HCH. In the early 1980s, levels of HCB were also considerably higher. They are close to the quantification limit now because of more stringent emission controls.

The Elbe is the main source of the pollutants in the German Bight. At the freshwater edge of the river the

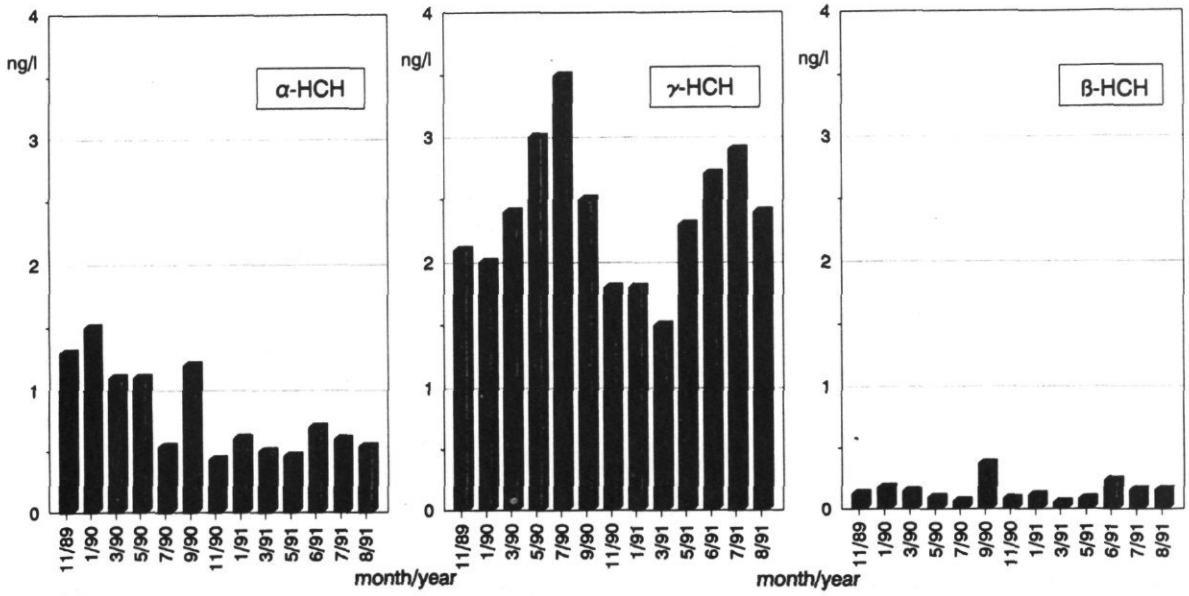


Figure 2. Seasonal variations of micropollutants. Station El 9, 54°00'N 008°05'E.

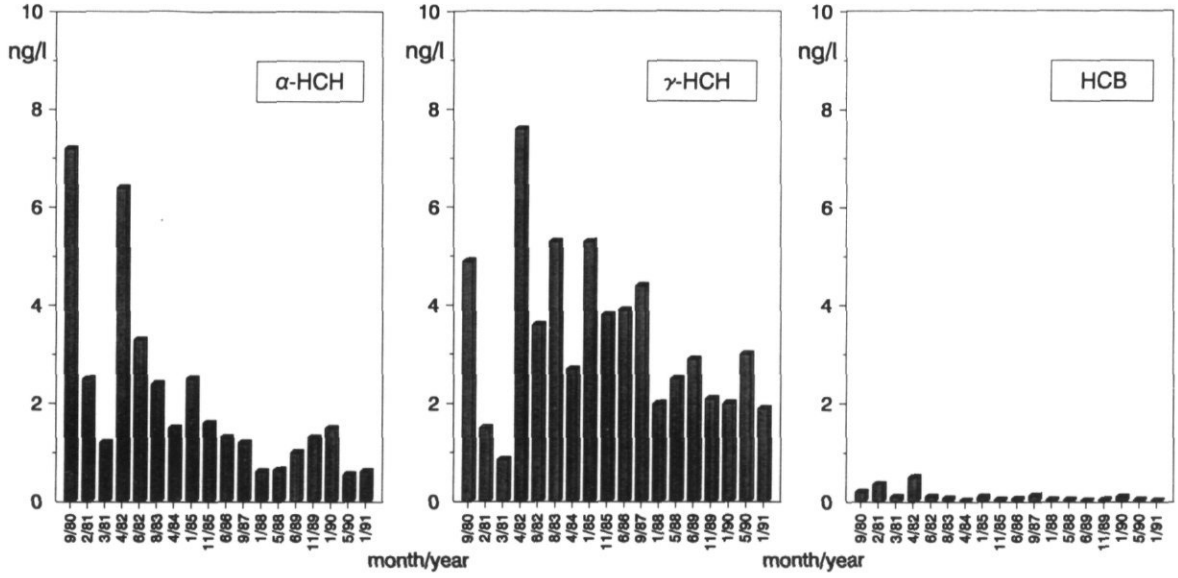


Figure 3. Temporal trend of organic micropollutants in JMP-Area 13. Station El 9, 54°00′N 008°05′E.

water is well mixed under tidal influence but still undiluted by saline water. Concentrations measured there can be computed with runoff volumes to given input data (Table 2) (ARGE Elbe, 1991). The data are average means of 8 to 12 data sets for each year. As much as a tenfold reduction for α-HCH and HCB can be seen as well as the variability of the "unchanged" input of lindane by a factor of about 2 between low and high data between 1981 and 1989.

Arkona Basin

In the Baltic Sea the "Arkona Basin" station has been sampled every two years for 15 years as part of a national monitoring programme. The station has now been declared a representative station (BMP-K4) of the Baltic Monitoring Programme (BMP) (Fig. 4). Because it is an offshore station, a comparison with input data of a nearby source is not possible.

In the surface water of BMP-K4 a 50% decrease of α-HCH can be observed. For lindane, no downward trend is obvious. HCB concentrations of BMP-K4 were close to or below the quantification limit, so no clear statement can be made.

Spatial trend

North Sea

The spatial variability in the North Sea is quite considerable. The contamination pattern is produced by the location and strength of sources of input, the flushing conditions, and by the residual current that governs the route of transportation.

About 90% of the contaminated freshwater input stems from the large rivers draining into 10% of the North Sea area to form the continental coastal water in the Southern and German Bight. The residual current transports this input counter-clockwise and parallel to the shore into the Skagerrak, where it mixes with the outflow from the Baltic Sea. The Norwegian coastal current then removes the combined plumes of contamination from the North Sea. In the NW North Sea, the rapid exchange with Atlantic water keeps the concentration low, near baseline values.

This general lindane pattern in 1981 was confirmed five years later, although levels had greatly increased (Fig. 5) (DHI, 1988). The distribution patterns of α-HCH and HCB are shown in Figures 6 and 7, which depict similar general features.

Table 2. Input from the Elbe in tons/year (ARGE Elbe 1991) mean values (n = 8–12).

	1981	1982	1983	1984	1985	1986	1987	1988	1989	1990
α-HCH	0.8	0.5	0.4	0.1	0.3	0.1	0.2	0.1	0.13	0.12
γ-HCH	0.8	0.5	0.7	0.6	0.7	0.5	0.8	0.5	0.44	0.21
HCB	1.2	0.5	0.3	0.1	0.2	0.1	0.05	0.08	0.11	0.08

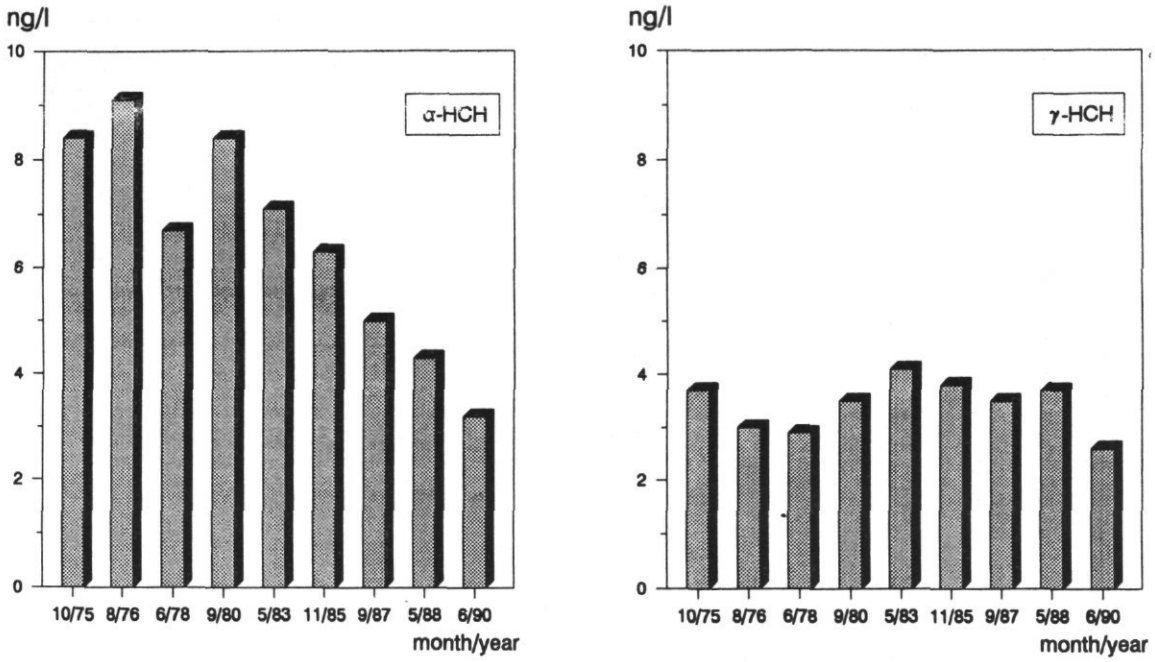


Figure 4. Concentrations of HCH's in surface water of the Baltic Sea. Arkona Basin BMP-K4, 55°00'N 14°05'E

Baltic Sea

In the Baltic Proper, the levels of some contaminants are higher because of the low water exchange. The Skagerrak, Kattegat, and Belt Sea are exchange and mixing zones between the North Sea and the Baltic Sea (Gaul, 1984). In the Skagerrak, higher concentrations of pollutants are found in the outflow of surface water, while near-Atlantic water concentrations are found in the inflow of bottom water (Fig. 8).

In the Kattegat and Belt Sea, concentrations approximate and become homogeneous east of the Darss Sill with small differences between the various basins that form the Baltic Sea.

The significant decline of α-HCH already highlighted in the Arkona Basin time series (Fig. 4) is now demonstrated for the entire Baltic Sea (Fig. 9). The time series did not reveal a trend for lindane. The southward shift of the 2.0 and 3.0 ng/l isopleth, however, could indicate how the compound is slowly rinsed out of the Baltic (Fig. 10).

Northeast Atlantic

Throughout the Northeast Atlantic, only α- and γ-HCH are found only in background concentrations apparently caused by atmospheric deposition. Elevated concentrations of both compounds are found in the Norwegian current, which is contaminated by the combined outflows from the North Sea and the Baltic Sea (Gaul, 1992).

Baseline concentrations for α- and γ-HCH were investigated in the Norwegian Sea in 1979 and 1985. The levels there have not changed much, but there are signs of a decrease. The average means, however, overlap in their variances and the difference is still insignificant.

1979 α-HCH = 1.32 ± 0.34 ng/l (n = 14)
γ-HCH = 0.33 ± 0.07 ng/l (n = 14)

1985 α-HCH = 1.03 ± 0.25 ng/l (n = 28)
γ-HCH = 0.22 ± 0.08 ng/l (n = 28)

(Figs. 11 and 12).

Conclusion

The spatial distribution pattern of contaminants is influenced by:

- the location and strength of sources of input
- the rate of degradation or entrapment
- the water exchange
- the residual current
- the level of background values.

The recognition of a temporal trend depends on:

- the rate of the trend
- the length of the time series available
- the natural variability of the determinant
- the analytical error of the method for the determinant.

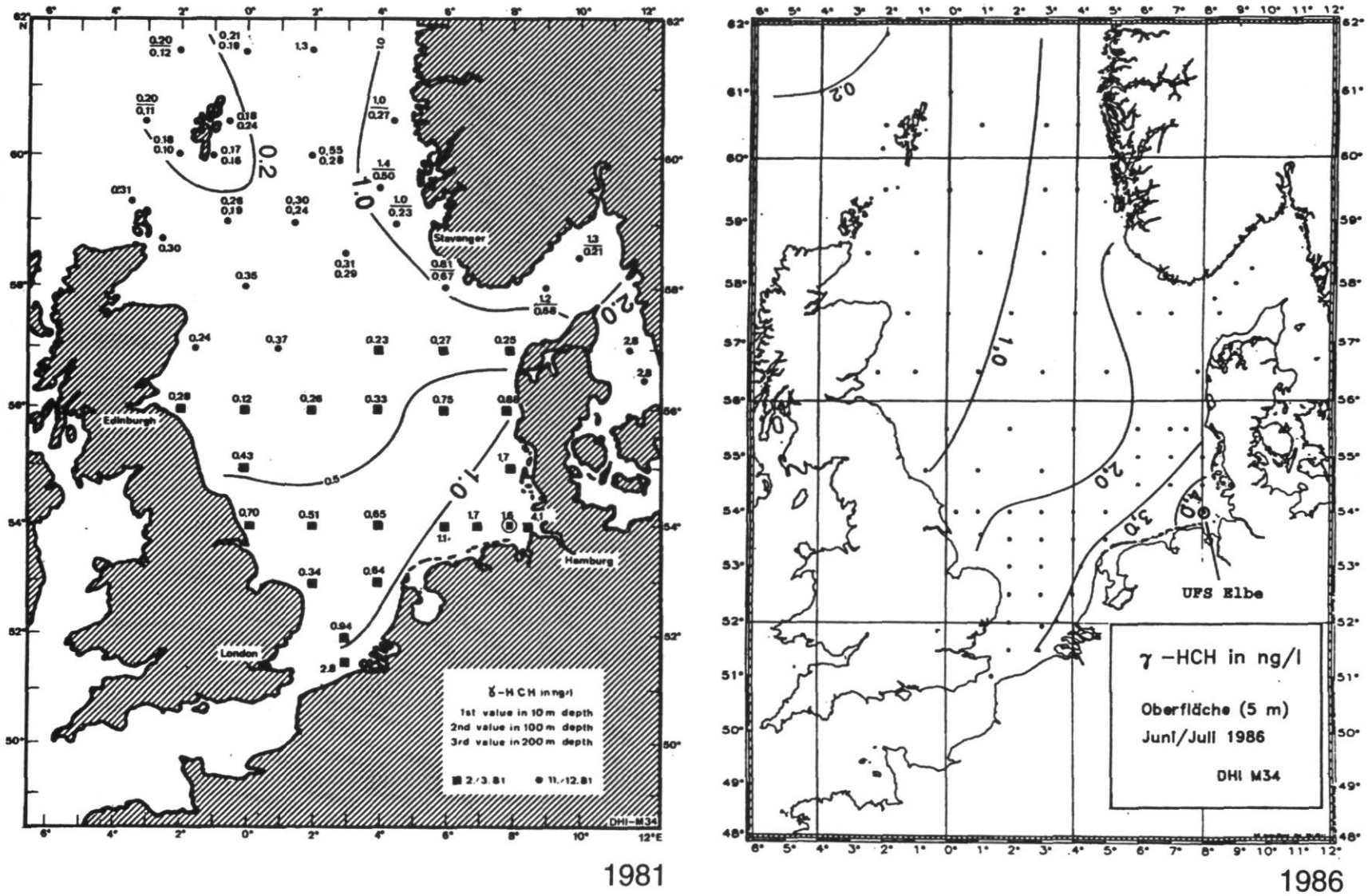


Figure 5. Surface water concentration of lindane for 1981 and 1986.

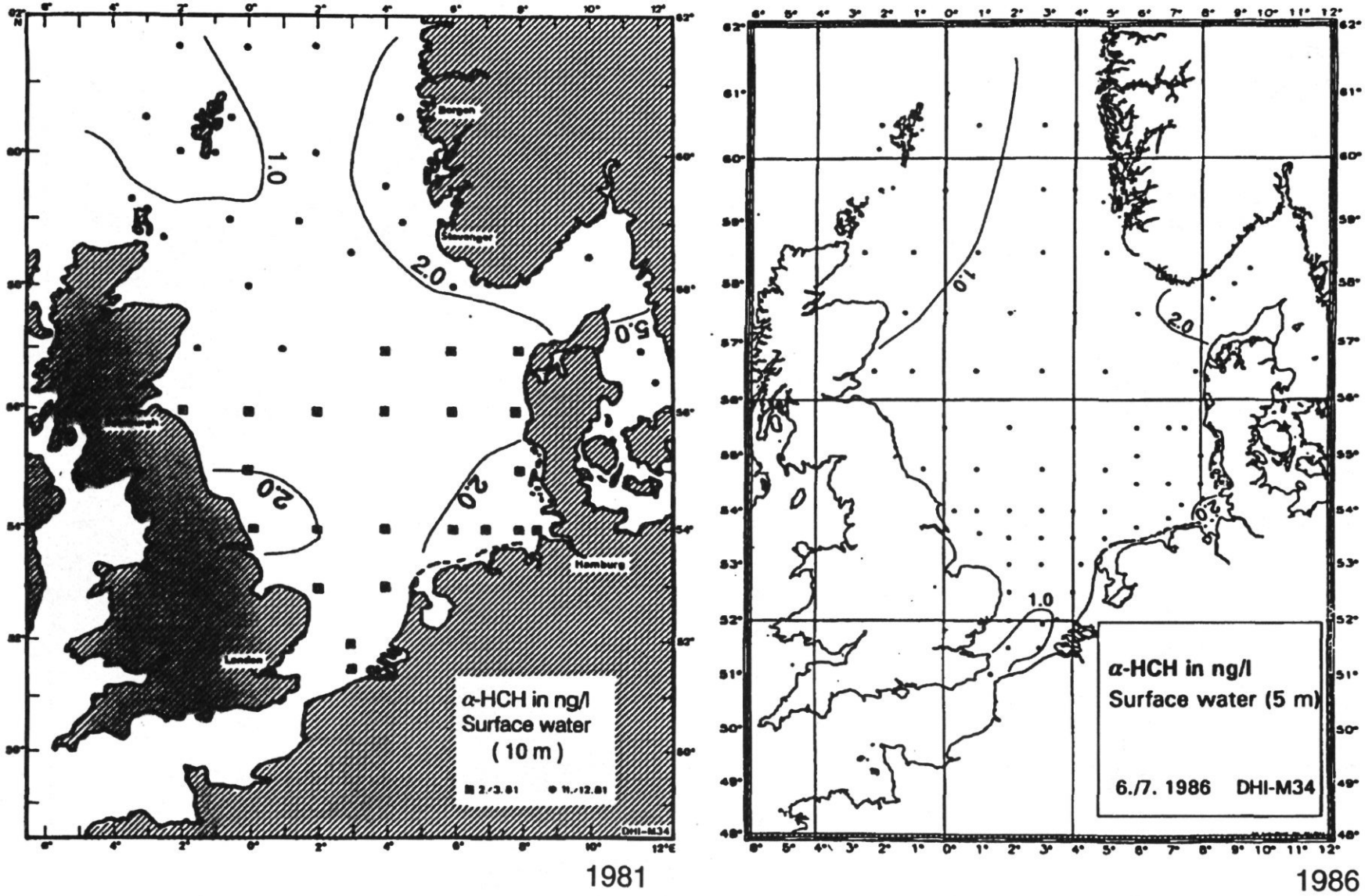


Figure 6. Surface water concentration of α-HCH for 1981 and 1986.

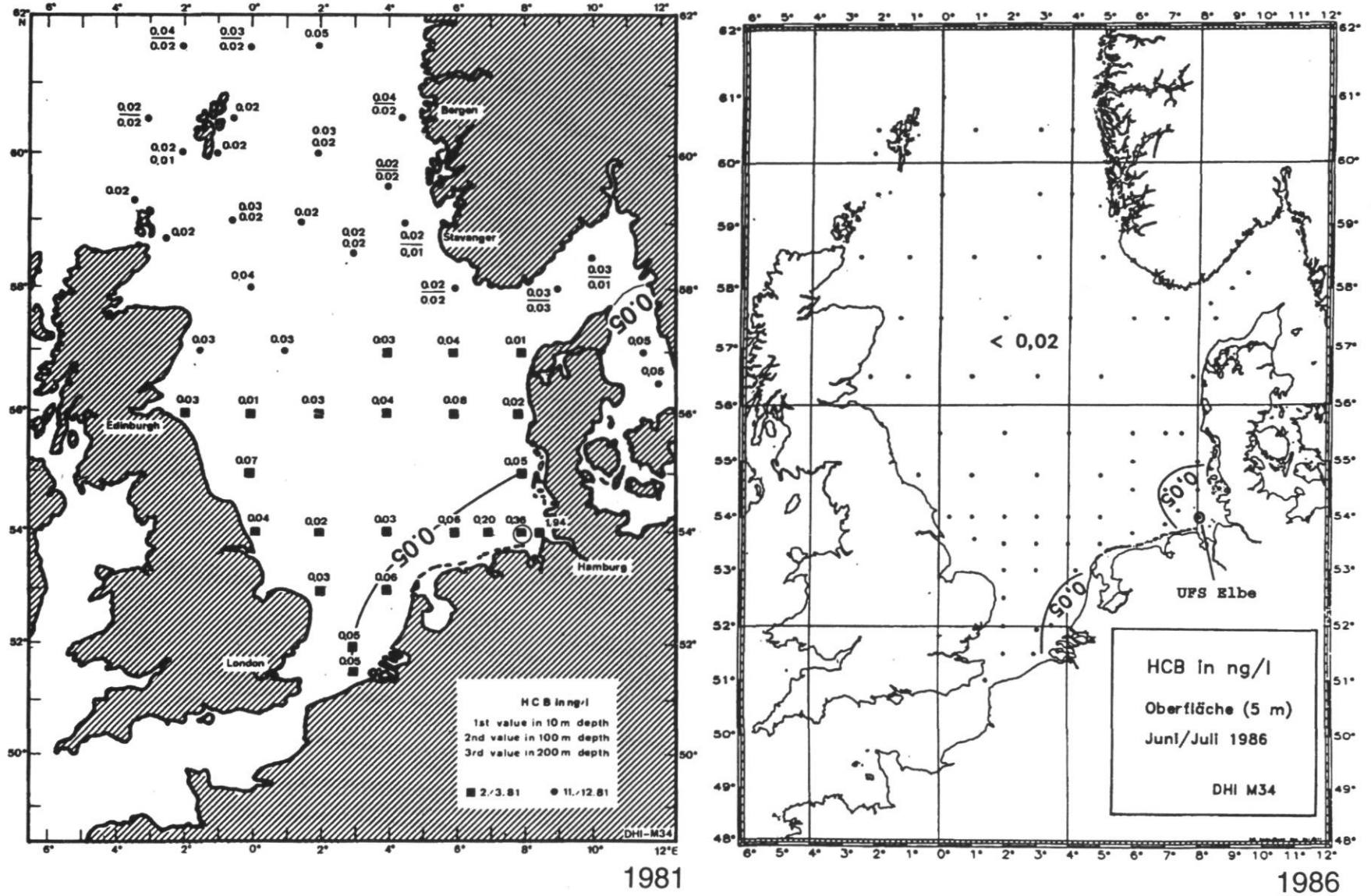


Figure 7. Surface water concentration of HCB for 1981 and 1986.

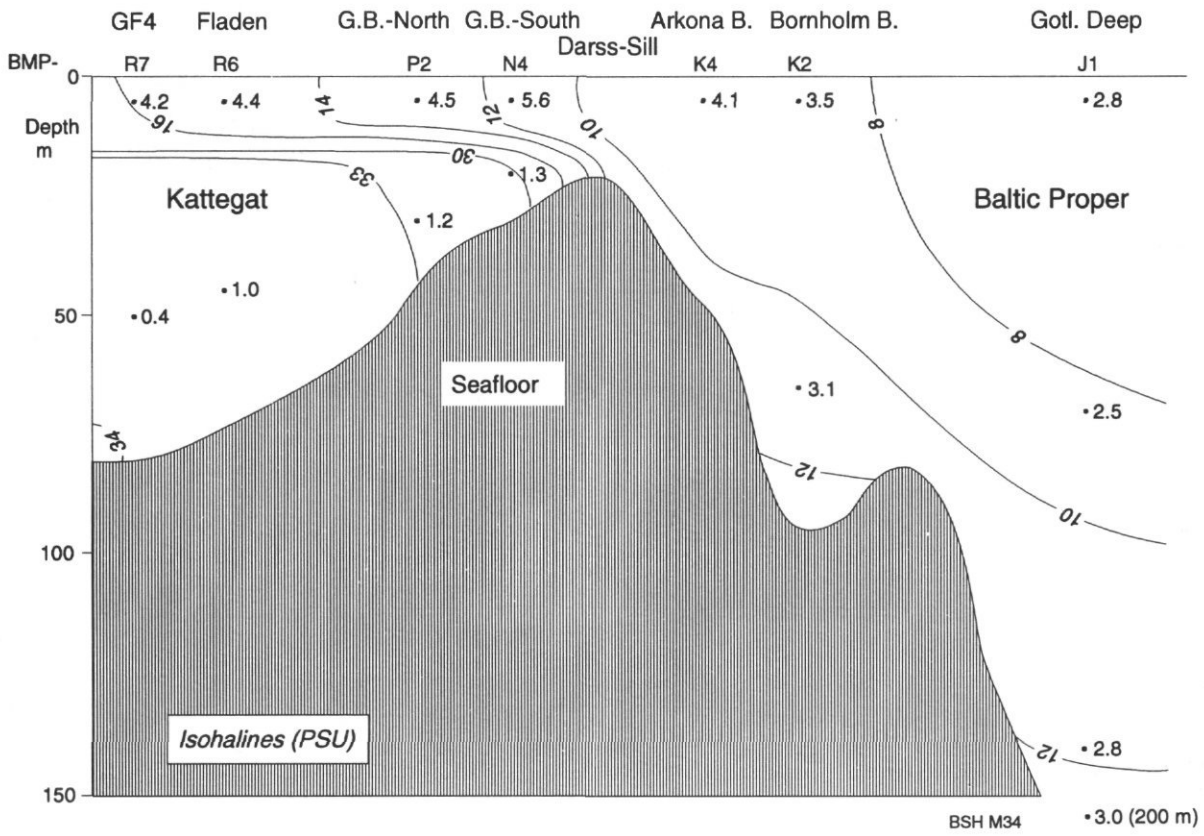


Figure 8. Lindane from Kattegat to Gotland Deep in ng/l (5/6/1983).

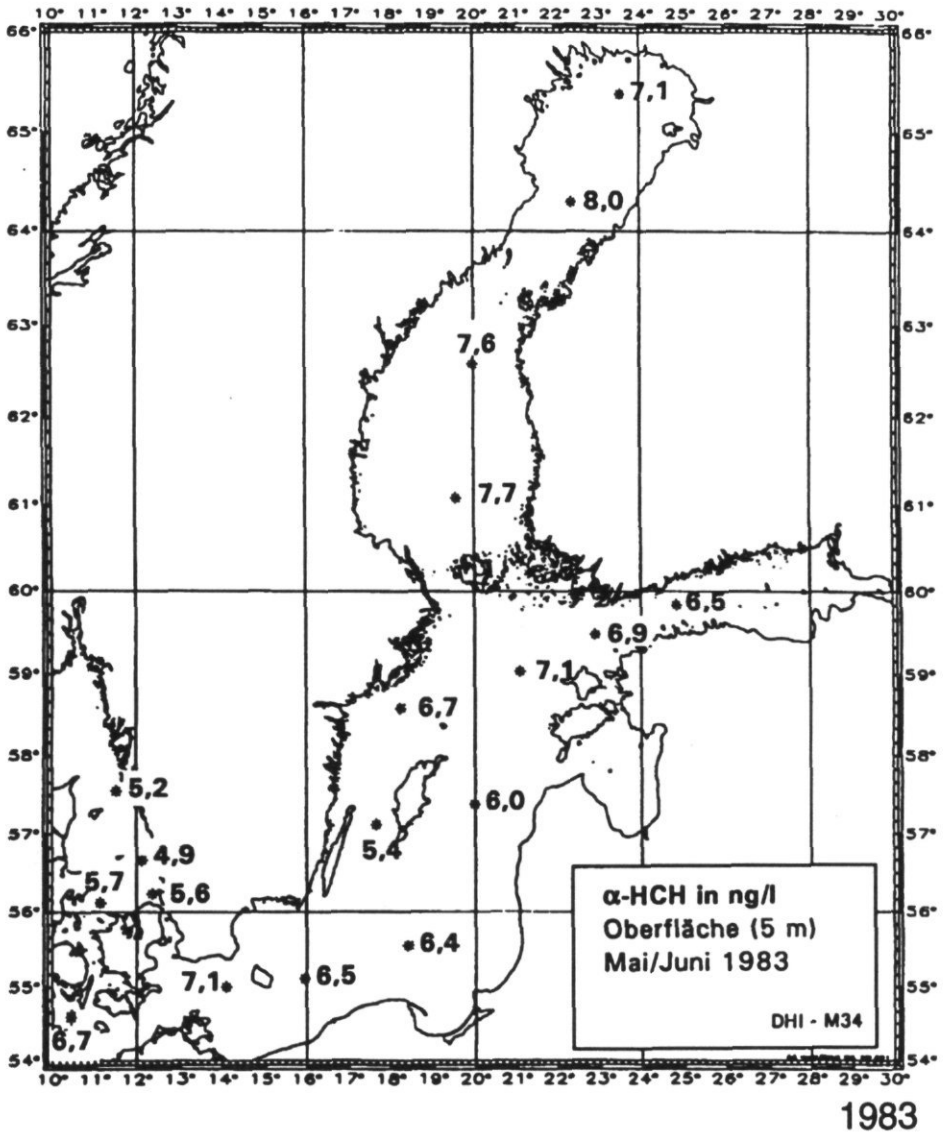
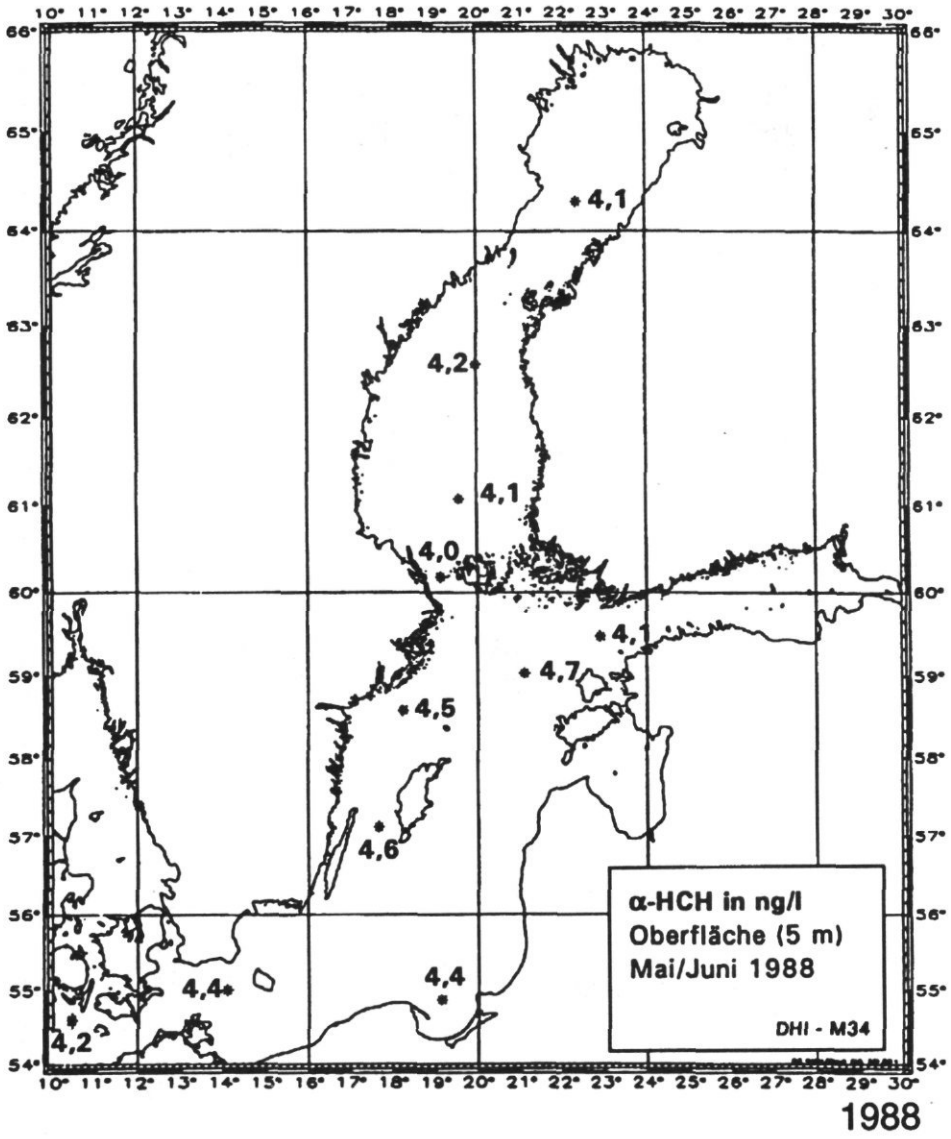


Figure 9. Surface water concentration of α-HCH (1983 *above*, 1988 *facing*).


4,1
4,2
4,1
4,0
4,1
4,7
4,5
4,6
4,4
4,4
4,2
α-HCH in ng/l
Oberfläche (5 m)
Mai/Juni 1988
DHI - M34

1988

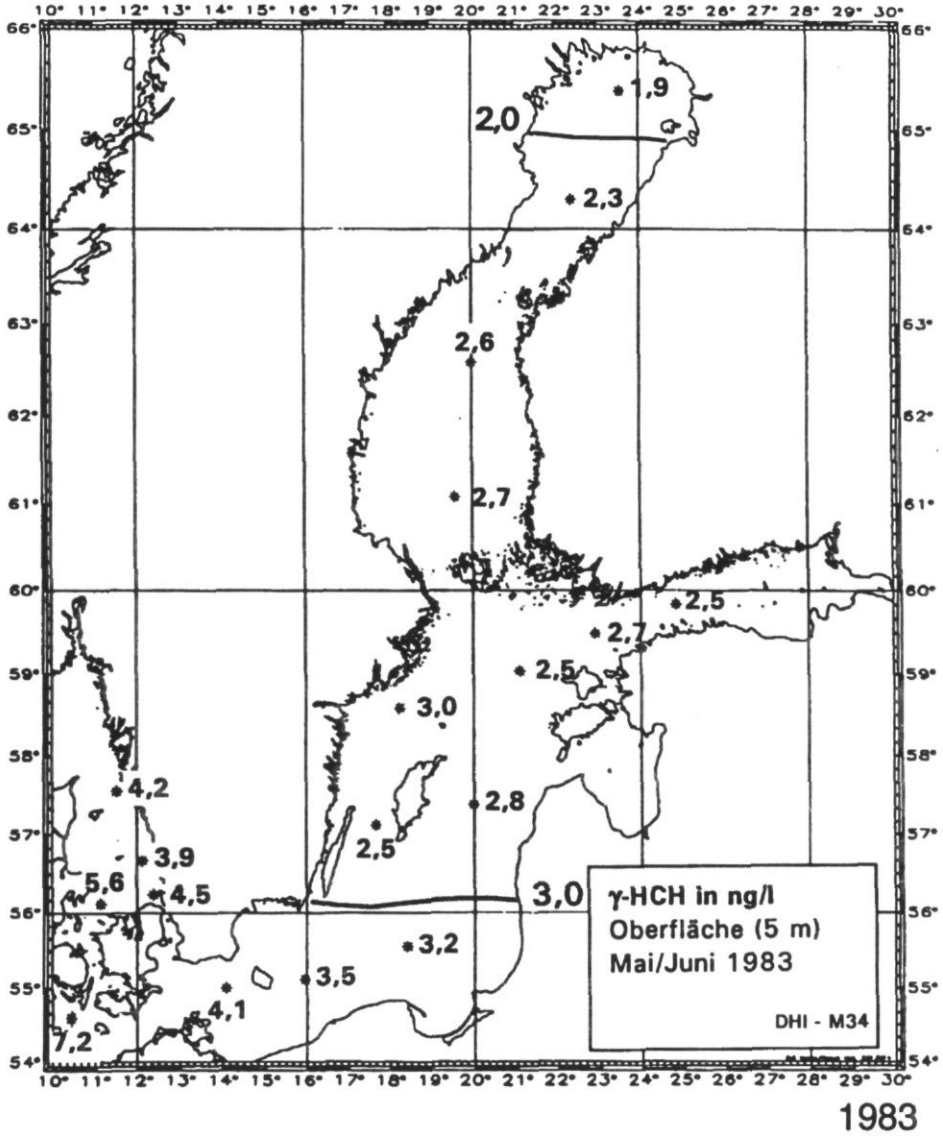
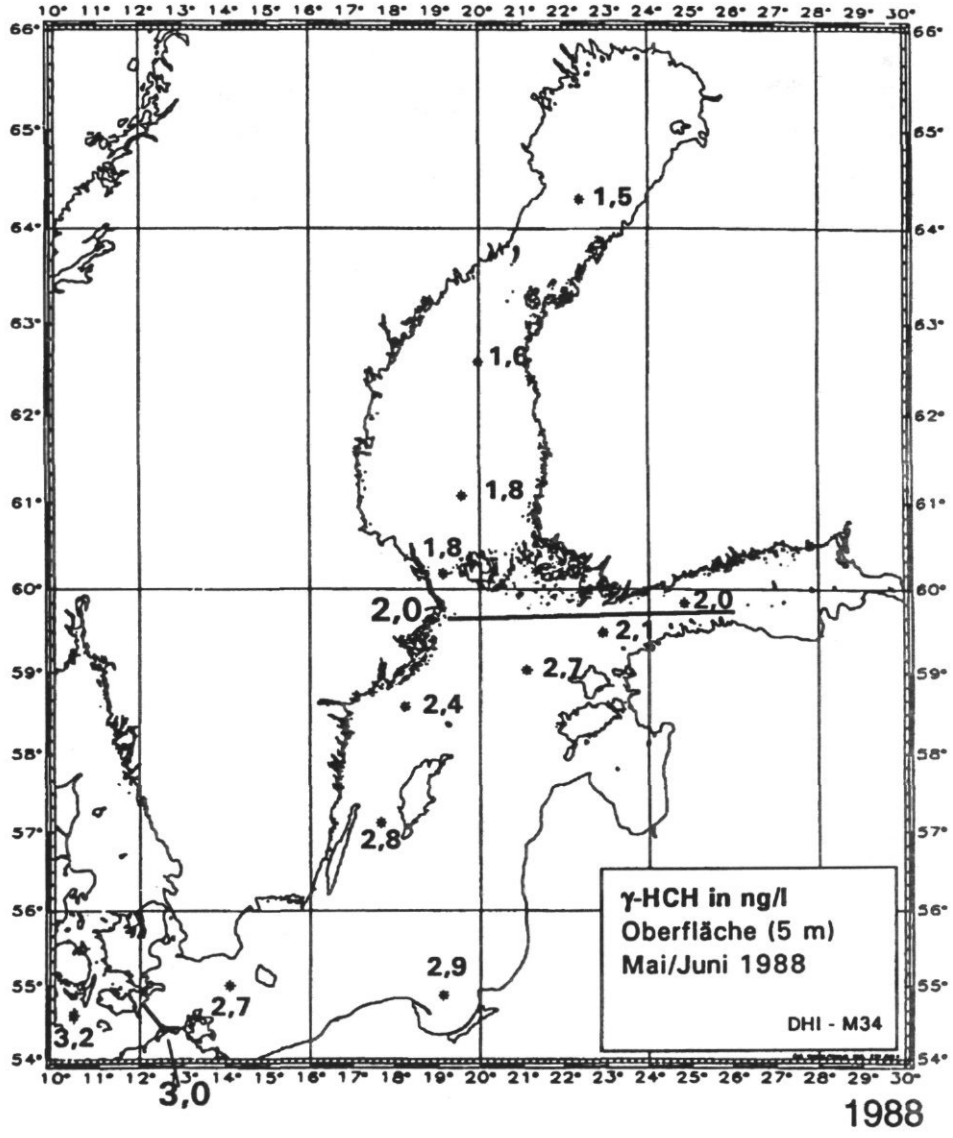


Figure 10. Surface water concentration of γ-HCH (1983 *above*, 1988 *facing*).


1,5
1,6
1,8
1,8
2,0
2,0
2,1
2,7
2,4
2,8
2,9
2,7
3,2
3,0
γ-HCH in ng/l
Oberfläche (5 m)
Mai/Juni 1988
DHI - M34
1988

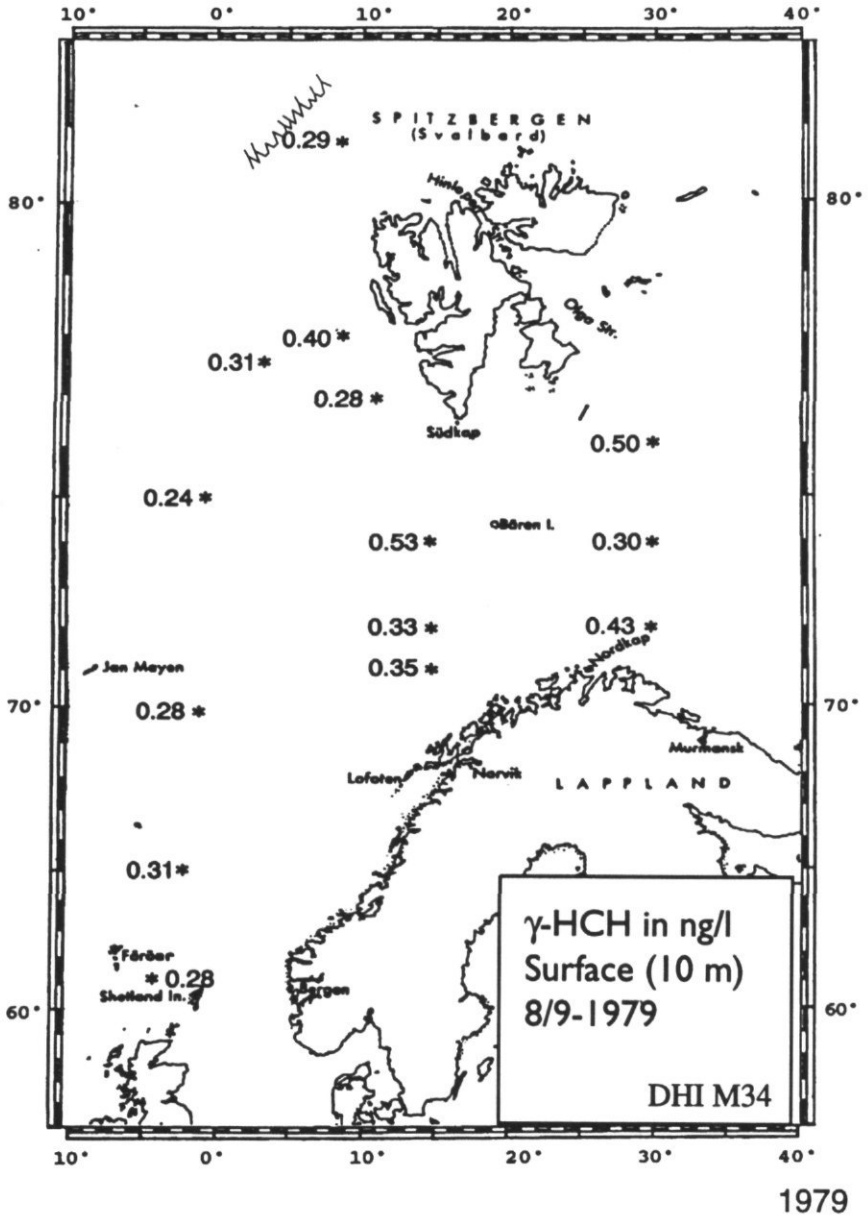
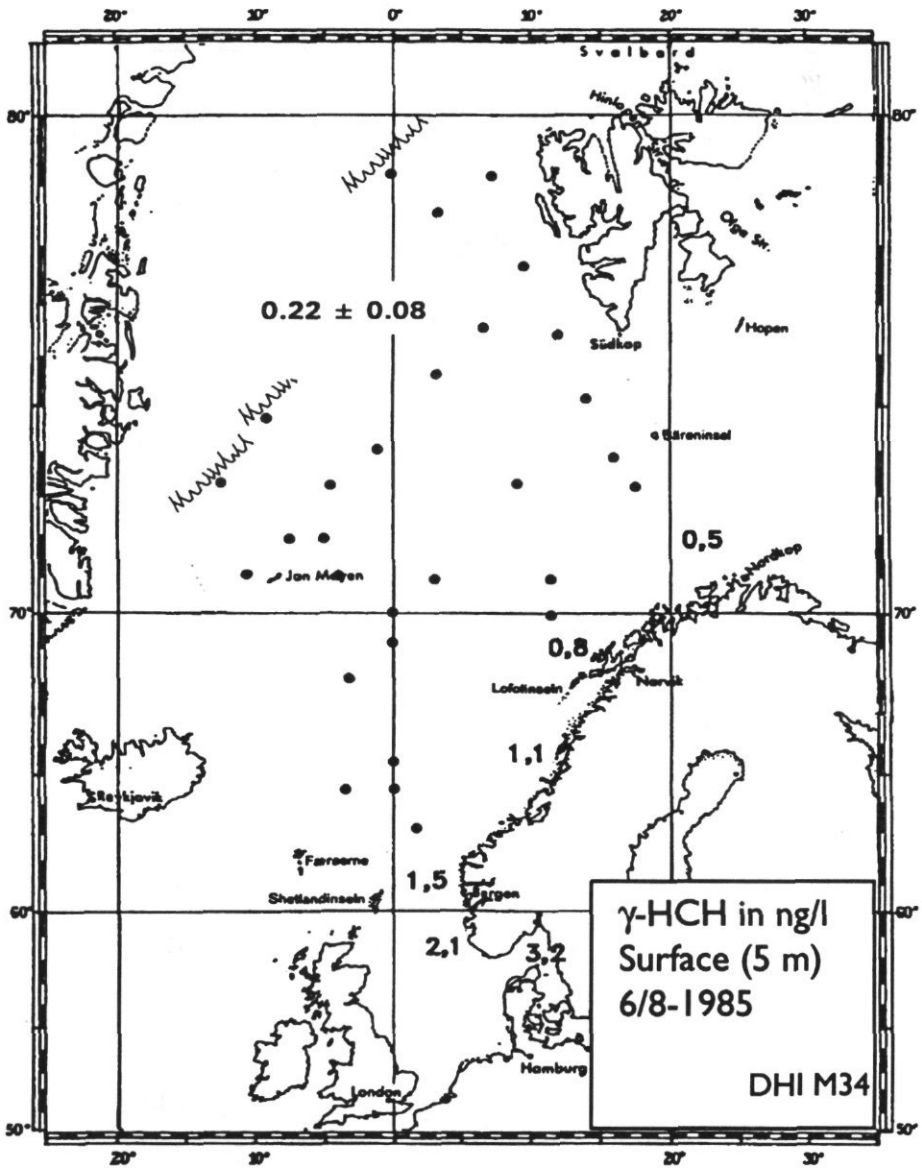


Figure 11. Surface water concentration of lindane (1979 *above*, 1985 *facing*).


Svalbard
0.22 ± 0.08
0,5
0,8
1,1
1,5
2,1
3,2
γ-HCH in ng/l
Surface (5 m)
6/8-1985
DHI M34

1985

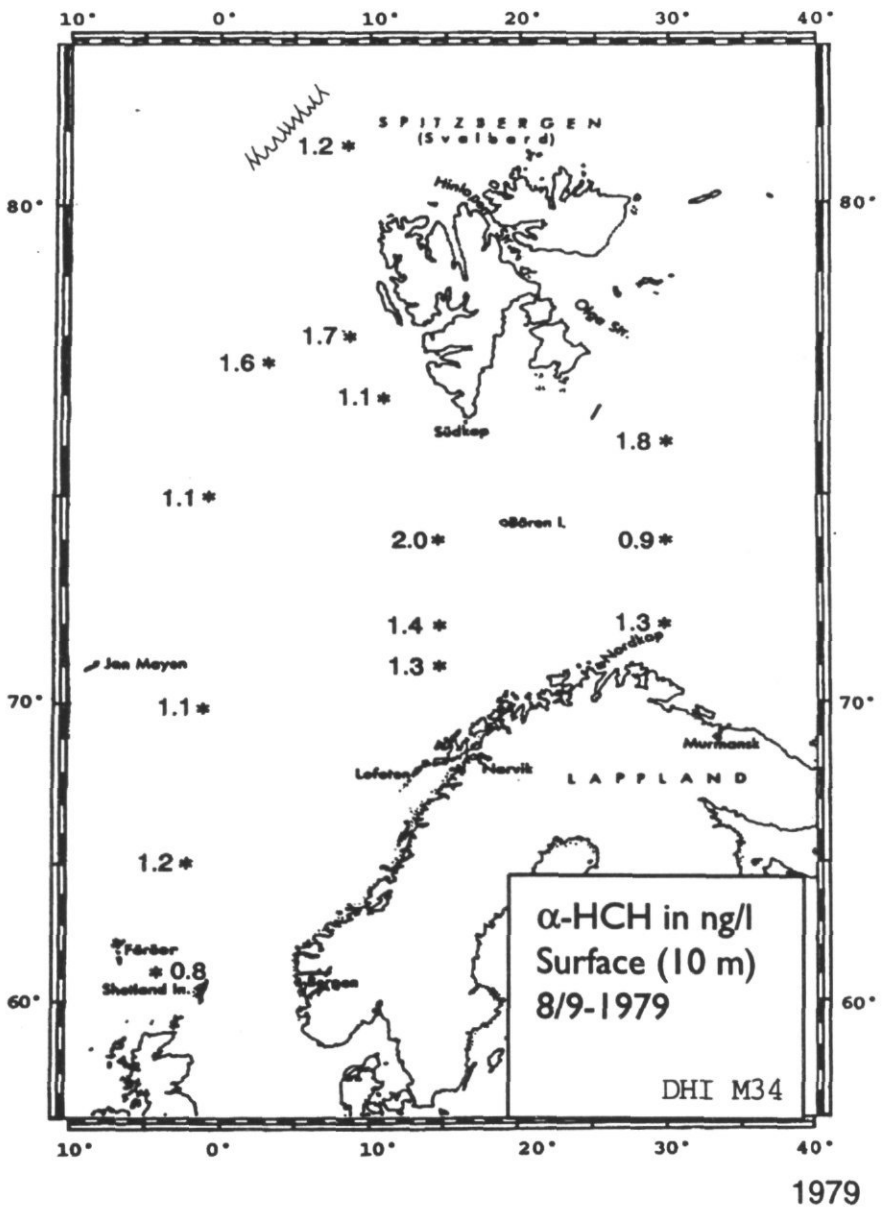
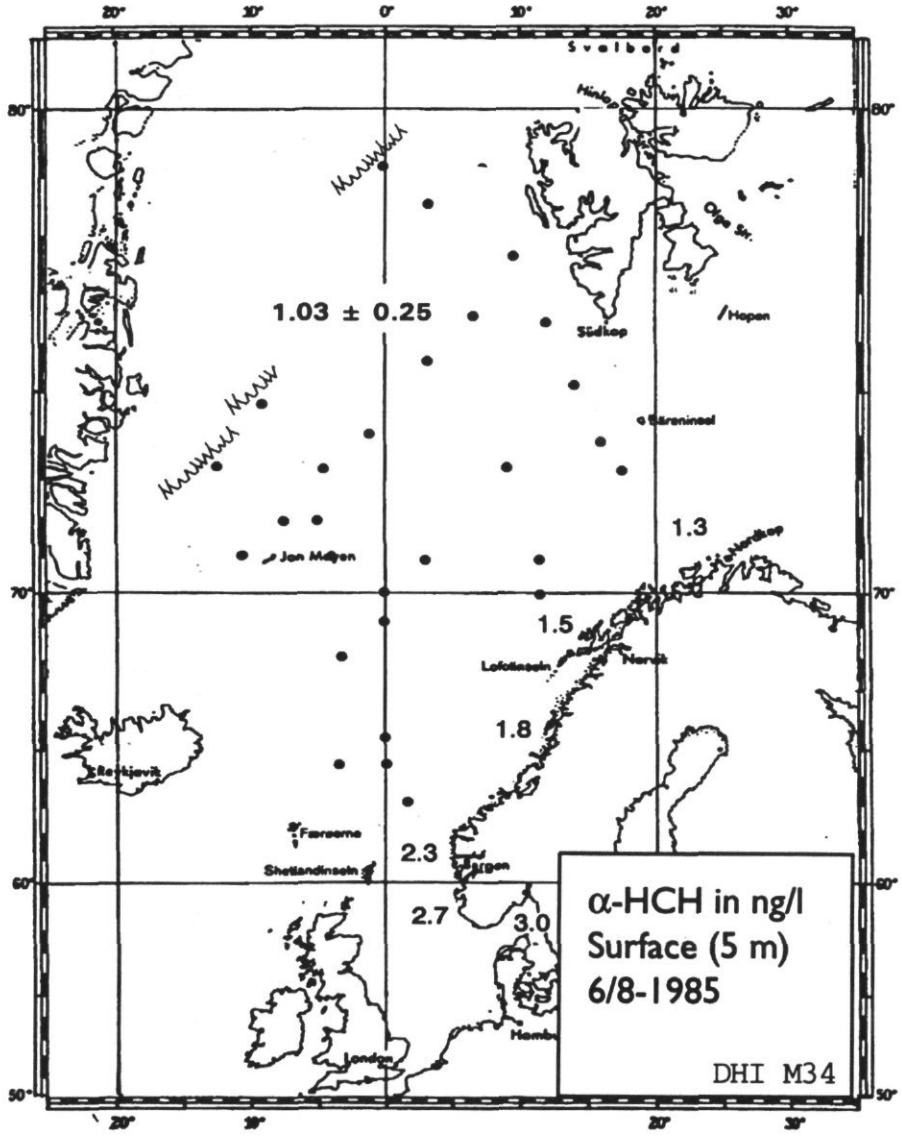


Figure 12. Surface water concentration of α-HCH (1979 *above*, 1985 *facing*).



1985

Spatial or temporal trends can best be studied using xenobiotics like organochlorines, as their background values and natural variability are low.

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ICES mar. Sci. Symp., 195: 127–143. 1992

The East Shetland Atlantic Inflow

W. R. Turrell

Turrell, W. R. 1992. The East Shetland Atlantic Inflow. – ICES mar. Sci. Symp., 195: 127–143.

Since oceanographic research began in the northern North Sea in the late 19th century debate has continued on the major routes of entry of Atlantic water into this semi-enclosed shelf sea. Early evidence was often ambiguous and prone to errors of interpretation, but direct current meter observations in the 1970s suggested a major inflow occurred from the west between the Orkney and Shetland Islands; the Fair Isle current (Dooley, 1974). This work using current meter observations also suggested that no significant, coherent inflow occurred further north, to the east of Shetland, contradicting many earlier studies employing indirect evidence. Recent direct measurements, obtained during the summer months when the northern North Sea was predominantly vertically stratified, have suggested that a more northern Atlantic inflow does exist, or did during the late 1980s. The measurements revealed a persistent current lying offshore from the Fair Isle current in the northwestern North Sea. It has been directly observed during two consecutive autumn periods, and has been identified as a salinity maximum in the bottom waters of the northern North Sea. This feature has been observed at the same geographical location in four consecutive years. Cross-correlations showed it was independent of the Fair Isle current, and had a different response to wind forcing. A volume budget of the northern North Sea suggested that the inflow entered the North Sea to the east of the Shetland Islands. Hydrographic observations east of Shetland suggest that the current there was narrow and was confined to a 30-km wide coastal strip, at least during the summer months. Seasonal variation in the western inflows to the northern North Sea is discussed using simple linear models of wind and density-driven transports. Limited biological evidence is also suggestive of a long-term change in the circulation of the northern North Sea, with reduced Atlantic inflow occurring during the mid to late 1970s. Some hypotheses relating to the possible physical mechanisms responsible for these changes are discussed.

W. R. Turrell: SOAFD Marine Laboratory, PO Box 101, Victoria Road, Aberdeen, AB9 8DB, Scotland.

Introduction

Charts showing the circulation patterns within an area of the ocean often have a remarkable influence on the evolution of concepts and theories relating to that area. These charts are always, at best, an individual author's concept of the mean circulation within an area, built up from many pieces of fragmentary evidence collected in different subareas at different times. They rarely convey the degree of short-term variability that may be imposed upon the mean circulation. Nor do they convey the possibility of long-term, interannual change that may occur with alterations in the physical driving forces as a result of climate change. For example, certain features of circulation charts derived from observations are often chosen to confirm the results of complex numerical models (while other features not reproduced by the model are just as often ignored). Biological oceanographers are often forced to resort to the simple and rigid concepts of circulation derived from such charts in order to explain and interpret processes such as planktonic drift and observed fish distributions.

Thus, although circulation charts are still a valuable method of summarizing many observational results, providing a framework around which concepts and hypotheses may be constructed, they must be used with caution and consistently reviewed by experiments designed to examine short-term, seasonal, and interannual variability. Such a process is, of course, constantly applied to the North Sea by scientists from its many bordering countries. This paper reviews the results from one such international experiment conducted in the northern North Sea during 1987 and 1988, the Autumn Circulation Experiment (ACE), and presents more

recent data from experiments stimulated by ACE. Like its predecessors, this paper also arrives at a conceptual circulation chart which is now open to examination and review.

Historical background

Over the past 100 years oceanographers have debated the possible routes of entry of Atlantic water into the North Sea. Evidence has come from surface drift bottles and cards (e.g. Fulton, 1897; Tait, 1934), from temperature and salinity distributions (e.g. D'Arcy Thomson, 1909; Robertson, 1909), and from planktonic indicator species (e.g. Fraser, 1952). Although each of these techniques suffered from limitations when used to deduce the paths of residual advection resulting from transport by ocean currents, they generally suggested two possible major inflows from the west; one through the passage between Orkney and Shetland, the other around the north of Shetland, and then southwards into the northern North Sea.

Direct measurement of currents over long periods had to await the general introduction of recording current meters in the late 1960s. Dooley (1974) analysed some of the first results from current meters deployed on moorings within the northern North Sea. He proposed a circulation pattern with the principal western inflow to the North Sea entering as a current flowing between Orkney and Shetland (the Fair Isle current). His circulation pattern (Fig. 1b) excluded only persistent inflow from the north, to the east of Shetland. Water movements east of Shetland were thought to be weak and variable, dependent to a large extent on local wind forcing (Dooley, 1974).

Most of the experiments from which our current knowledge of North Sea dynamics is derived were carried out during the 1970s. They took place mainly during the winter period, when vertical stratification was absent. An example is the international JONSDAP '76 experiment (Joint North Sea Data Acquisition Program, 1976). Results from this study have proved most useful in calibrating the two-dimensional, vertically integrated numerical models which are used today for surge prediction and pollutant transport simulation (e.g. Davies and Heaps, 1980).

It was not until 1987 that a similar exercise was repeated during a period when the northern North Sea exhibited extensive seasonal vertical stratification. The international Autumn Circulation Experiment began in September 1987 and extended through the period when vertical stratification in the northern North Sea was eroded by autumn gales, until January 1988 when vertical mixing was virtually complete (Bartsch *et al.*, 1990;

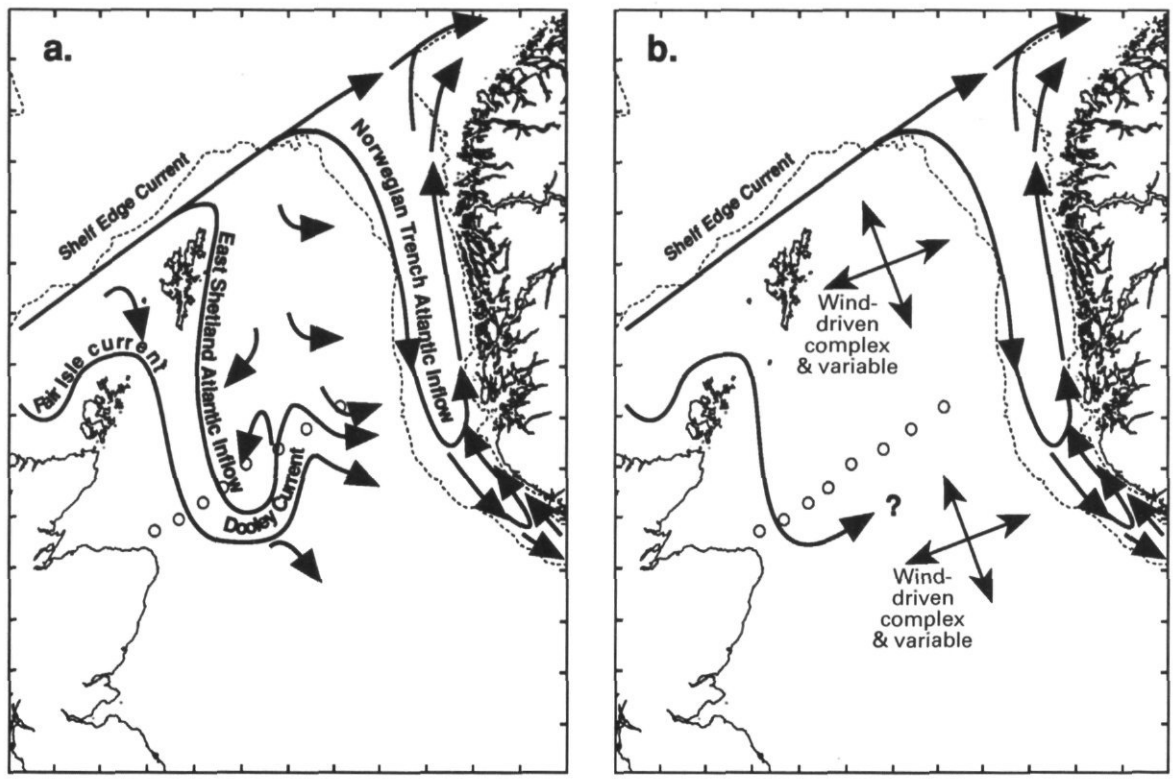

Figure 1. Charts showing the conceptual models of the circulation of the northern North Sea: (a) as derived by Turrell (1992) using the results of the Autumn Circulation Experiment (ACE); and (b) as derived by Dooley (1974).

Turrell *et al.*, 1990). During this experiment a wealth of physical evidence was collected which revived the notion that there was a persistent and well defined inflow into the North Sea from the north, independent of the Fair Isle inflow at least during the summer months (Turrell *et al.*, 1992; Svendsen *et al.*, 1991). The circulation scheme, shown in Figure 1a, resulting from this experiment has been partly confirmed by more recent work east of Shetland.

Recent evidence for the East Shetland Atlantic Inflow

Recording current meter observations

A line of eight moorings, the East Coast (EC) line, extending out from the Scottish east coast along 058°T into the northern North Sea was deployed during ACE (Fig. 2). Figure 3a shows a vertical section through the EC line with contours of monthly-mean current speeds, resolved perpendicularly to the section.

Moving out from the coast, a weak inflow occurred at the first, most westerly mooring, inshore from the steeper bottom slope found near the 100 m contour. A core of persistent inflow was observed at the second mooring, located above the 100 m contour, with a second area of weaker and more variable flow found at the current's offshore boundary, at mooring 3. This core flow, located above the 100 m contour, was thus thought to be the Fair Isle current observed across its southward leg (Fig. 1a and b).

A second persistent and stable inflow was observed at the fourth mooring (Fig. 3a). The monthly-mean vectors at the eight moorings are shown in Figure 3b. The two distinct areas of persistent inflow can be seen at moorings 2 and 4, separated by the area of weak and variable inflow at mooring 3.

To check the current meter results obtained during ACE, moorings 2, 3, 4, and 5 were redeployed during the autumn of 1988. Again two cores of inflow were observed, one at mooring 2 and the other at mooring 4 (Fig. 4). Thus the inflow observed at mooring 4 is a semi-

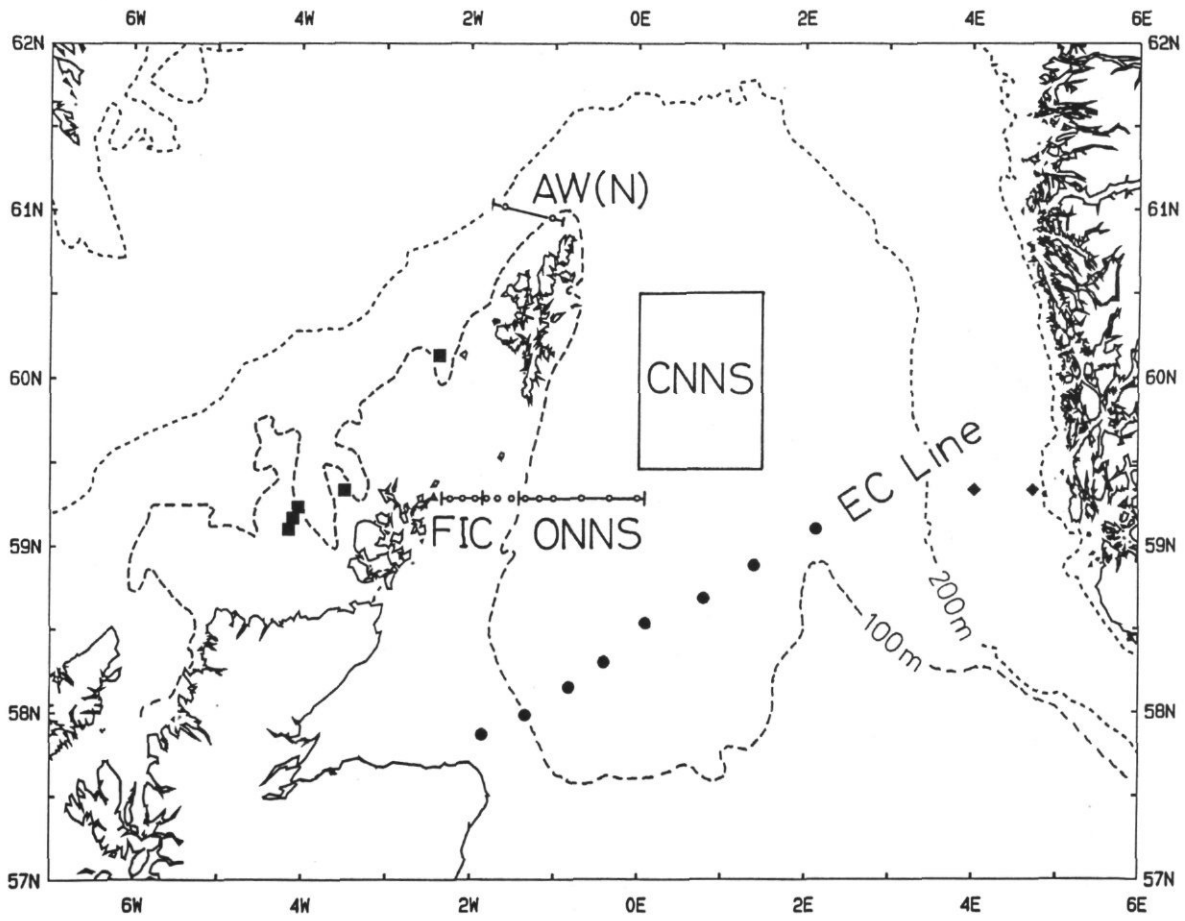

Figure 2. Map showing the location of the eight recording current meter moorings (solid circles) deployed by the Marine Laboratory, Aberdeen along the East Coast (EC) line during the Autumn Circulation Experiment. The subareas and sections used to compute the indices of annual change are also shown. The EC box is bounded by 57°N, 60°N, 2°W, and 2°E.

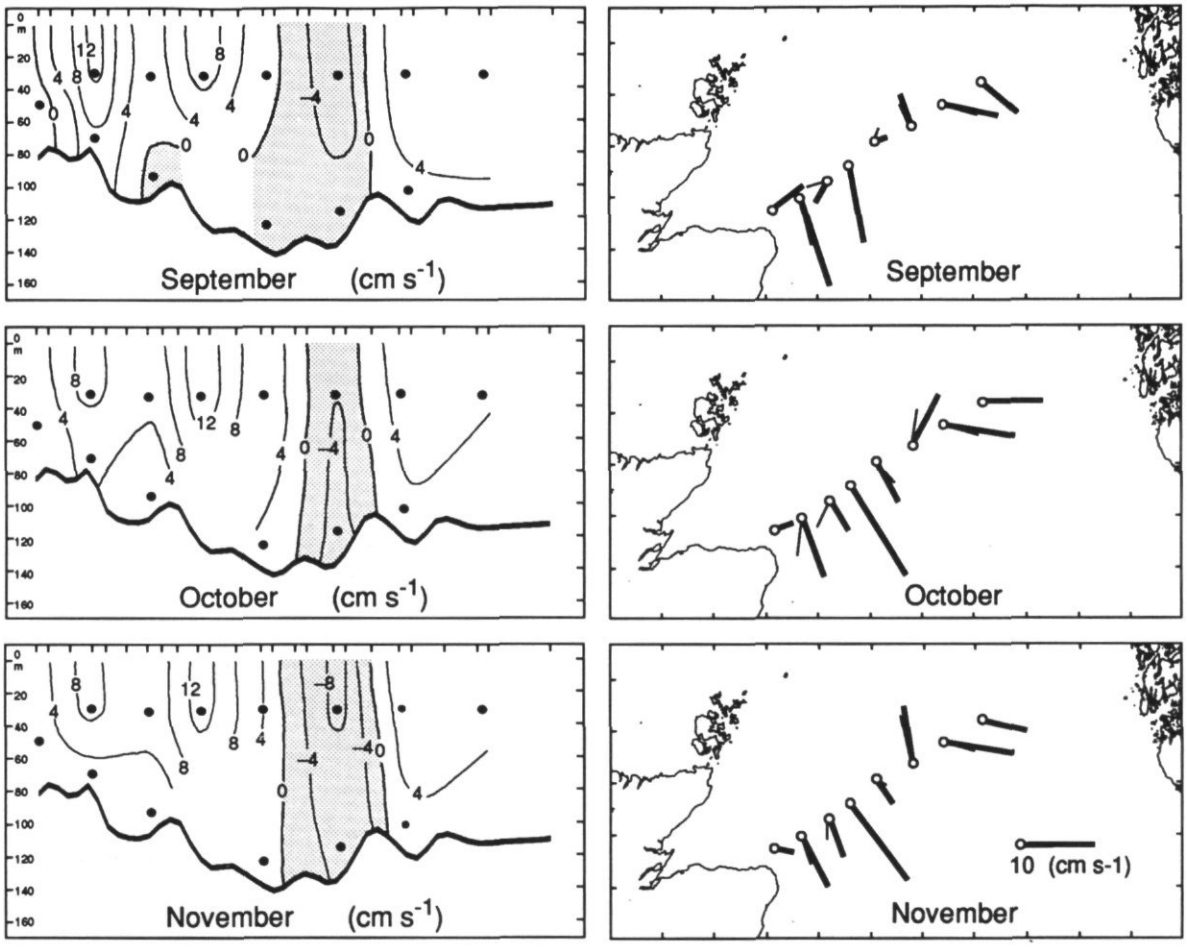

Figure 3. Representations of the monthly mean residual currents recorded at the eight current meter moorings deployed along the EC line during ACE (1987). Results are presented for the months when vertical stratification was evident in the northern North Sea. (a) Left-hand figures are contoured values of current speed resolved perpendicularly to the EC line. (b) Right-hand figures are mean current vectors along the EC line (thick lines are near-surface records, thin lines are near-bed).

persistent one, having been directly observed during two autumn periods. The observations show that the current flow was fixed in location, implying that it was steered by the topography to a certain extent.

The areas of outflow observed during ACE (Fig. 3) represent the eastern edge of a cyclonic eddy circulating around the deeper region of the northern North Sea, near the Fladen Ground (Svendsen *et al.*, 1991). The outflow is centred close to mooring 6 along the EC line. East of this mooring, at moorings 7 and 8, a third area of inflow occurs, and may be associated with the edge of the persistent inflow which occurs along the western slope of the Norwegian Trench. This third Atlantic inflow to the North Sea has been well documented in the past (e.g. Furnes *et al.*, 1986).

Temperature-salinity characteristics

In order to determine the origin of the inflows observed along the EC line of current meter moorings, the temperatures and salinities measured during September 1987 were deemed to be characteristic of the period when the northern North Sea was vertically stratified. The vertical section of temperature and salinity along the EC line is shown in Figure 5. The temperature–salinity diagram constructed from values from below the seasonal thermocline ($d > 60$ m) is shown in Figure 6.

The variable inflow at mooring 1 lay within warm ($T \sim 11.5°C$) and relatively fresh ($S \sim 34.90$) coastal water. Water within the persistent current at mooring 2 had the typical summer characteristics of the Fair Isle current ($T \sim 10$–$11°C$, $S \sim 35.0$–35.15). These values were consistent with those observed during the same period within the origins of the Fair Isle current to the west of Orkney. The values result from the mixing of pure Atlantic water with Scottish coastal water (Turrell, 1992).

Cooler ($T \sim 8°C$) and more saline ($S \sim 35.23$) water was found at moorings 3 and 4. This water had characteristics similar to those observed in Atlantic water lying at

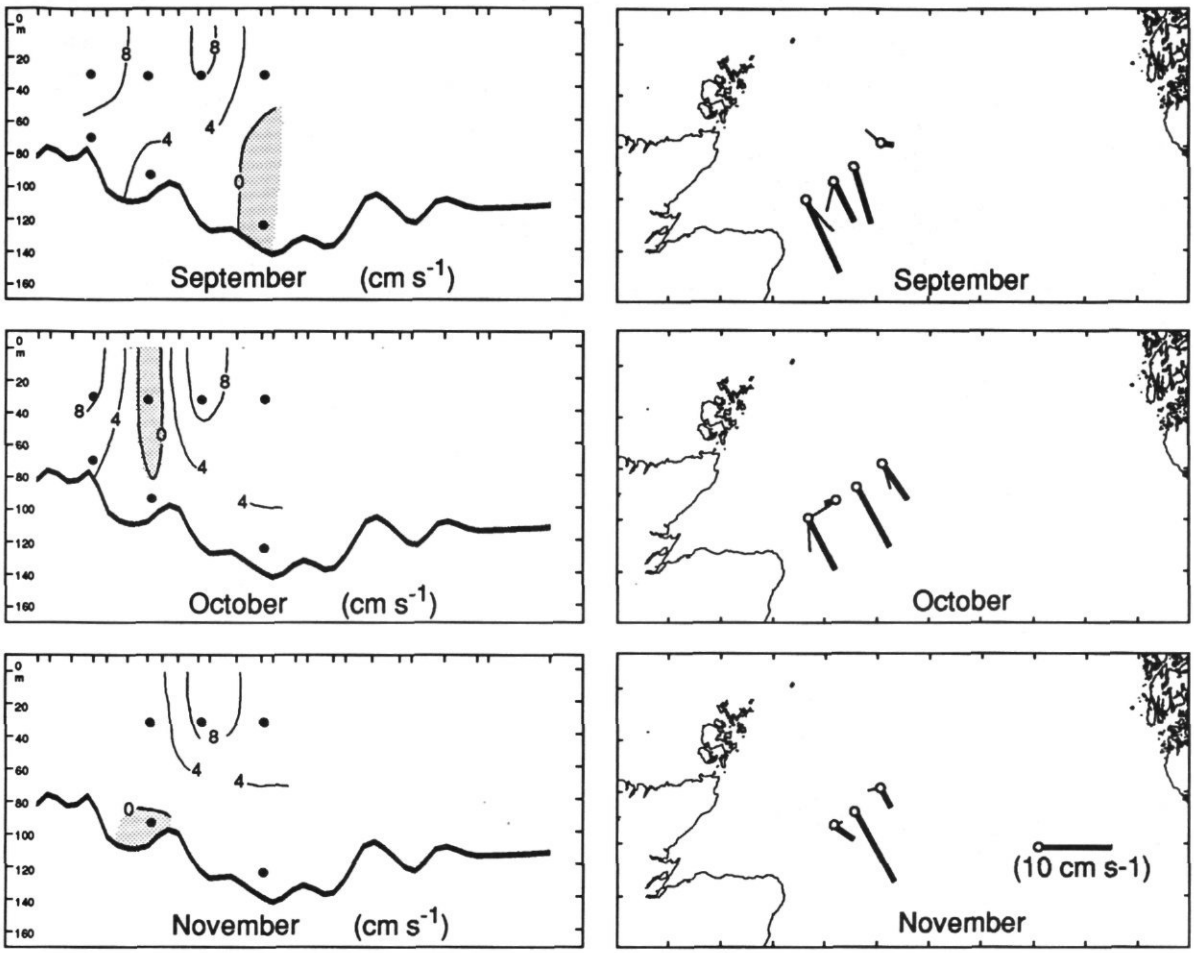

Figure 4. Results from the redeployment of current meter moorings 2, 3, 4, and 5 along the EC line during the autumn of 1988. Left-hand figures are contoured values of current speed resolved perpendicularly to the EC line. Right-hand figures are mean current vectors along the EC line (thick lines are near-surface records, thin lines are near-bed).

the shelf break to the north of the Shetland Islands. Hence it was assumed that water within the current observed at mooring 4 was of recent Atlantic origin and was not associated with the Fair Isle current itself.

The outflow at mooring 6 and the inflow at mooring 7 both occurred within water that was much cooler than either the Atlantic or coastal waters, with T ~ 6.5°C. This water was also fresher than the Atlantic water (S ~ 35.10) and is believed to be water that was formed in the northern North Sea during the previous winter mixing of surface waters, whose salinity had been reduced by the westward extension of coastal water from the Skagerrak and Norwegian coasts, with Atlantic water located within the North Sea. This water type has been named Cooled Atlantic Water (CAW – Svendsen *et al.*, 1991). Moving eastwards from mooring 7 to mooring 8 the temperature and salinity again increase in value approaching those of pure Atlantic water, hence confirming the presence of a third Atlantic inflow at the edge of the Norwegian Trench.

A salinity maximum at the same location in the central northern North Sea has been observed during four consecutive summer surveys along the EC line (Fig. 6). The transition out from the Scottish coast is from coastal water, through water typical of the Fair Isle current (FIC water), eastwards into water of intermediate temperature (cooler than the FIC but warmer than the CAW) but of maximum salinity. The FIC water type was observed generally close to the 100 m depth contour (mooring 2) while the salinity maximum indicating water of recent Atlantic origin was consistently found offshore from the Fair Isle current, at mooring 4.

Cross-correlations along the EC line

Additional confirmation that the current observed at mooring 4 in 1987 was independent of the Fair Isle current was obtained from cross-correlations between the seasonal low-pass filtered current meter records from the instruments along the EC line, and between

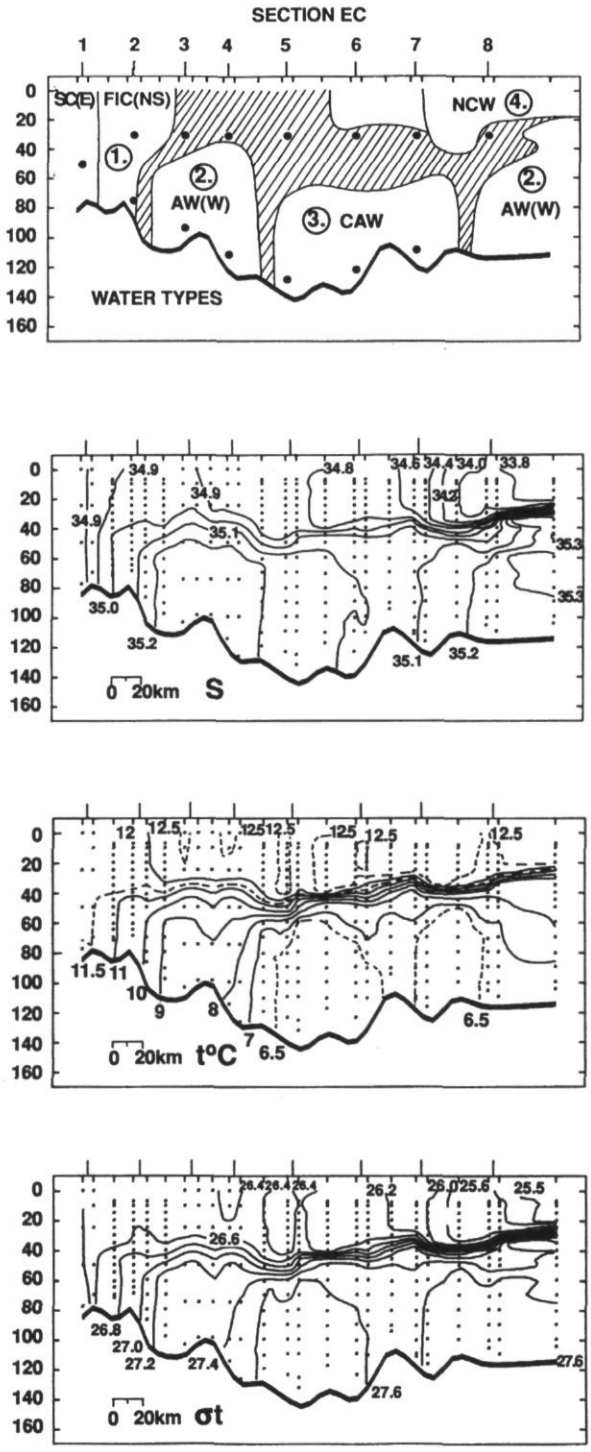

Figure 5. Vertical sections of temperature and salinity observed along the EC line in September 1987. The upper figure indicates the water types inferred from an analysis of the temperature-salinity characteristics. AW – Atlantic Water, FIC(NS) – Fair Isle current water as observed in the North Sea after the mixing that occurs during its entrance through the Orkney–Shetland Passage, CAW – Cooled Atlantic Water, NCW – Norwegian Coastal Water.

these records and a similarly filtered record of wind speed and direction obtained from Kirkwall, Orkney (Turrell *et al.*, 1992). The correlations between the current meter records selected for the months when the North Sea was stratified revealed three groups with significant correlations. The first included those records from both instruments at mooring 2 and the near-surface instrument at mooring 3. From the temperature and salinity observations, this group was assigned to the Fair Isle current. A second group occurred between records from the near-bed instrument at mooring 6, both instruments at mooring 7 and the near-surface instrument at mooring 8. This group was assumed to confirm the coupled inflow/outflow system occurring within the Cooled Atlantic Water lying within the deeper regions of the northern North Sea.

The third group with significant correlations was that obtained from the single instrument at mooring 4 and both instruments at mooring 5. When used in conjunction with the observed salinity and temperature distribution this evidence was assumed to confirm that an independent current consisting of water of recent Atlantic origin lay offshore from the Fair Isle current.

When the current at mooring 4 was compared to the wind, the current was found to be almost independent of local wind speed and direction (Fig. 7). It was evident that the current continued to flow towards the southwest with speeds of ~10–20 cm s^{-1} despite winds directly opposed to it and of variable strength. Closer analysis of the 1987 results revealed a markedly different behaviour than that observed within the Fair Isle current (Turrell *et al.*, 1992). While the current at mooring 2 during the summer months, demonstrated increased flow into the North Sea for winds directed from the north, the variability of the current at mooring 4 was correlated with winds from the east but with time lags of ~30–40 h. It was proposed that this may have been the result of easterly winds adjusting the density structure of the northern North Sea and hence altering the baroclinic forcing of the Atlantic Inflow as observed at mooring 4.

Volume transport budget of the northern North Sea

Evidence from current meter and hydrographic measurements has therefore confirmed the presence of a persistent flow of Atlantic water in the central northern North Sea during the summer months, lying offshore from the Fair Isle current. The route of entry of this water into the North Sea, however, was not directly observed. A volume flux budget of the northern North Sea for the month of October 1987 was determined using all the current observations made during ACE, and employing mass and salt conservation assumptions (Turrell *et al.*, 1992). The small influx of Atlantic water calculated as flowing towards the North Sea to the west of Shetland was insufficient to account for the flux

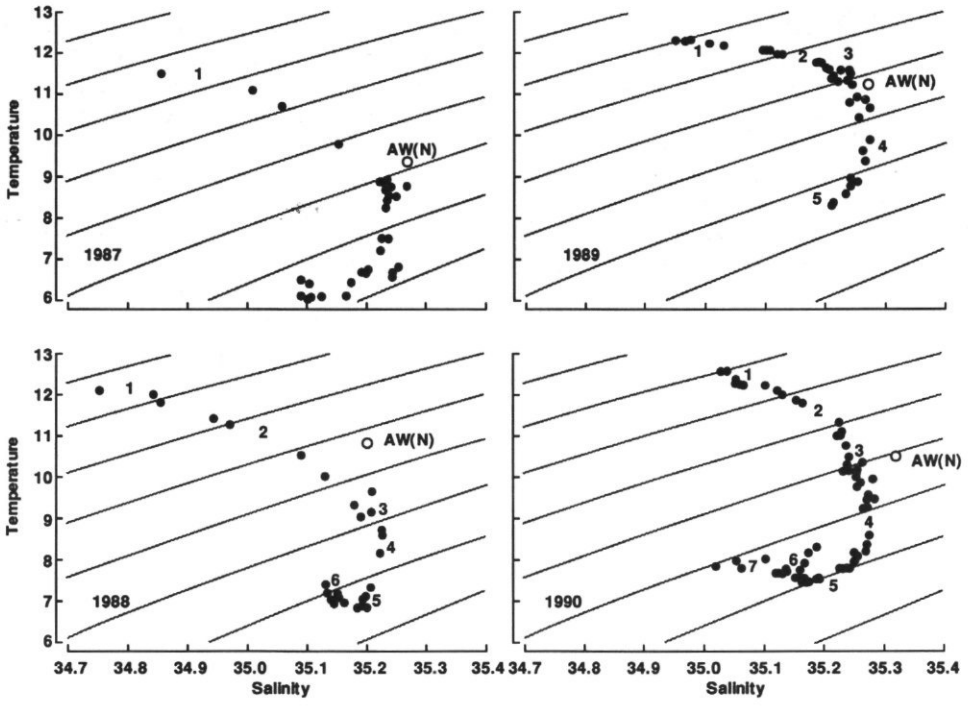

Figure 6. Temperature–salinity diagrams constructed using observations obtained along the EC line during September in four successive years. The numbers indicate the EC mooring locations in relation to the position of the CTD casts from which the data were derived. Data have been selected at 10 m intervals (in order to avoid clutter) from a depth of 60 m to the seabed (in order to avoid effects above the seasonal thermocline). The salinity maximum was always found close to the fourth mooring position.

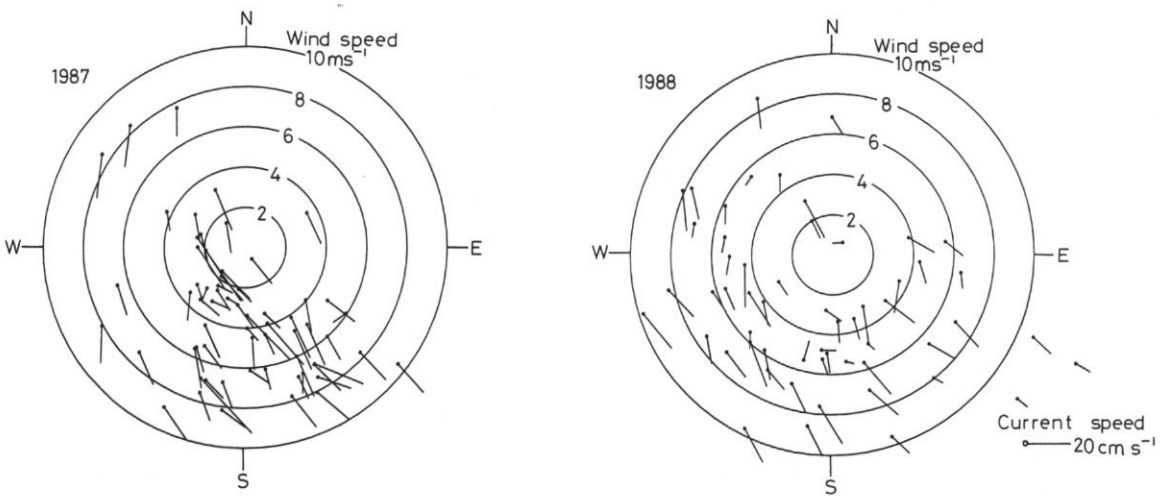

Figure 7. Wind–current hodograms demonstrating the persistence of the current of warm saline water found at the near-surface instrument (35 m) of mooring 4. Daily mean wind and current speeds and directions have been used. The location of the head of each day's mean current vector (small circles) indicates the speed and direction from which the wind was blowing. The direction away from the head in which the vector points indicates the direction to which the current was flowing, while its length indicates the speed of the current.

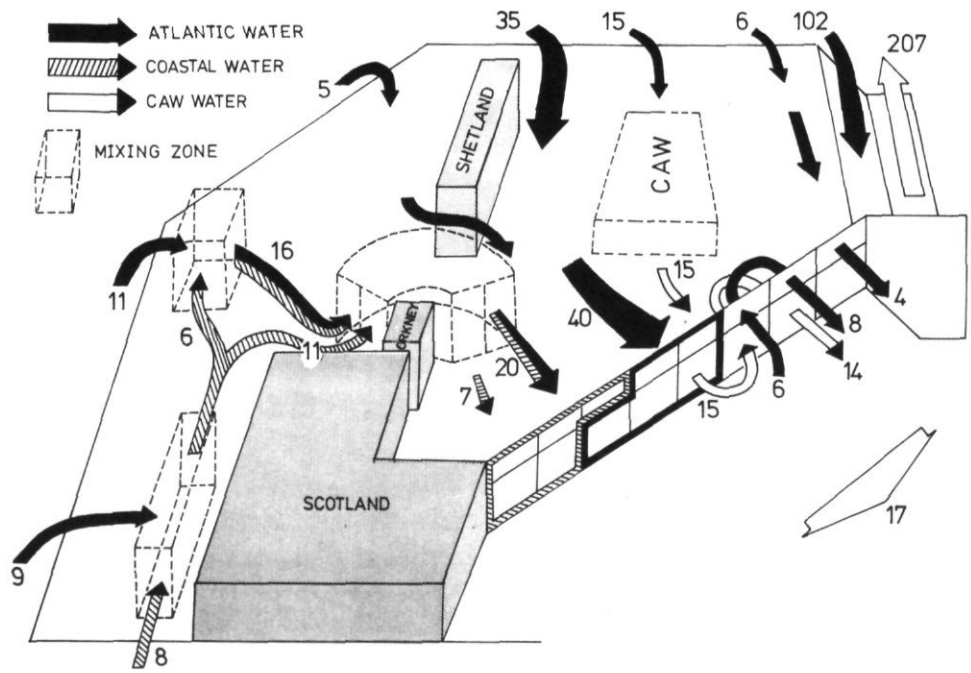

Figure 8. Schematic diagram showing the volume transport budget derived by Turrell *et al.* (1992) for the northern North Sea during October 1987.

observed at the EC line of current moorings. Hence it was postulated that the inflow occurred from the north, to the east of Shetland (Fig. 8).

Vertical mixing east of Shetland

The question as to whether the inflow was a broad diffuse one, as has been previously described for the northerly inflow to the North Sea (Otto, 1983) or a narrow confined current, was addressed during surveys carried out to the east of Shetland in 1988 and 1990 (Fig. 9). Hydrographic observations revealed a 20–30 km wide strip of well-mixed water extending down the east coast of the Shetland Islands. Numerical models of tidal mixing do not predict vertically well-mixed conditions in the deep water east of Shetland (Pingree and Griffiths, 1978) and hence this observation was taken to be indicative of a southerly flowing residual current advecting well-mixed water south from the intense mixing area around the northerly tip of the Shetland Islands.

Acoustic Doppler Current Profiler survey

A 25-h repeated survey within the strip of well-mixed water was completed in 1990 using a shipborne Acoustic Dopper Current Profiler (ADCP). A 15-km wide track was traversed 18 times over a 25 h period, and the resulting current vectors were resolved perpendicularly to the section and averaged over six subareas (Fig. 9). The resulting mean southerly velocities, when transformed into volume transports, represented a non-tidal inflow into the North Sea of approximately $26 \times 10^4\ m^3\ s^{-1}$ (Turrell, 1992). When extrapolation was performed to the seabed and across the entire width of the well-mixed zone it was determined that about $43 \times 10^4\ m^3\ s^{-1}$ of Atlantic water entered the North Sea through this narrow strip. This represented approximately 70% of the volume flux expected to enter from the north in order to arrive at a mass balance for the northern North Sea for the month of October 1987.

Seasonal variation of the inflows to the northern North Sea

The results presented above have been derived primarily from observations made during periods when the northern North Sea exhibited extensive vertical stratification. Observations made during the winter months following the autumnal erosion of the seasonal stratification were limited during ACE and subsequent current meter deployments owing to mooring losses. There were indications, however, that the inflow east of Shetland was reduced by a factor of at least two, while the Fair Isle inflow remained at the same magnitude, or had even increased.

In order to examine the seasonal characteristics of the two western inflows to the northern North Sea, the physical driving mechanisms responsible for their transports are now discussed and their annual variability examined. The simple transport models derived below have been used to arrive at a description of an annual

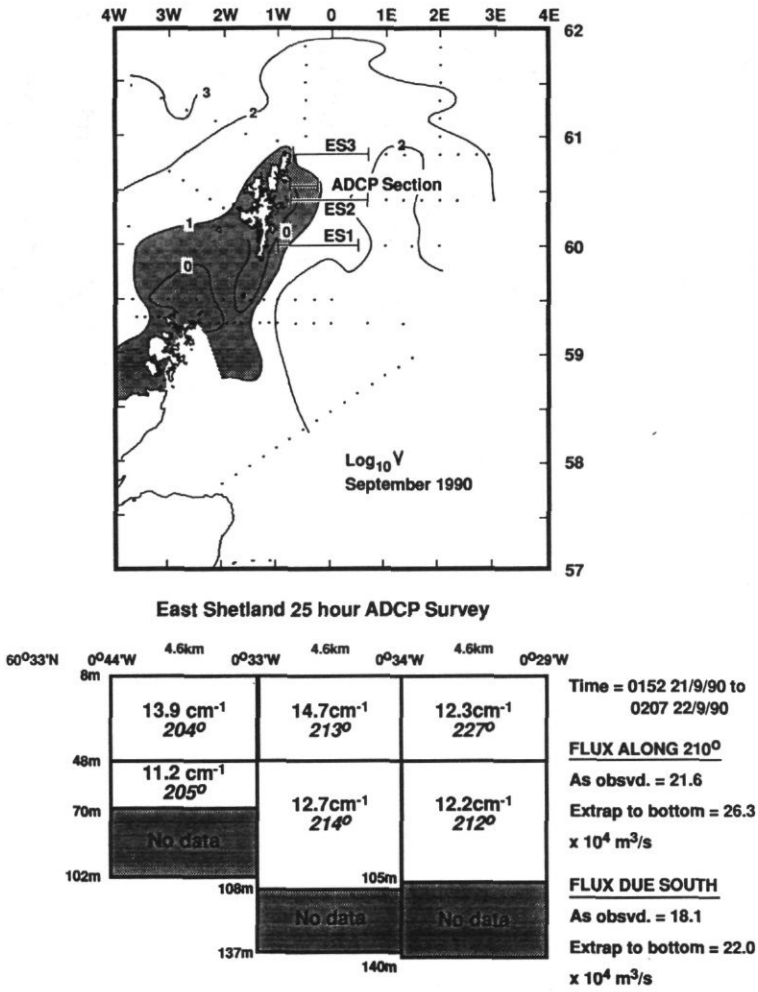

Figure 9. Horizontal contour chart of the stratification parameter ($\log_{10}V$ where V = potential energy anomaly). Values <1 are indicative of well-mixed conditions. The narrow strip of well-mixed water along the east coast of the Shetland Islands is not predicted by numerical models of tidal mixing (Pingree and Griffiths, 1978) and may be the result of advection southwards away from the area of intense mixing north of Shetland. The results of a 25-h ADCP survey within the well-mixed strip are also shown.

circulation cycle for the northern North Sea in the absence of a comprehensive series of direct observations.

Fair Isle current – west of Orkney

From observations across the Fair Isle current to the west of the Orkney Islands before it entered the North Sea, Turrell *et al.* (1990) deduced that the current was predominantly wind-driven during both winter and summer. They found a linear response between volume transport towards the North Sea and wind stress when this was directed from the south and southwest, while wind stress directed from the north and northeast resulted in outflow from the North Sea. The results confirmed those derived by Pingree and Griffiths (1980) from a two-dimensional numerical model of the wind-driven circulation in the northwest European shelf seas. They derived a simple regression equation describing the linear response of volume transport within the Fair Isle current (Q_{FIC}) to wind stress (τ_w) from varying directions:

$$Q_{FIC} = c + m \cos(\Theta - \Theta_{max})\tau_w \quad (1)$$

where:

Q_{FIC} = Volume transport within Fair Isle Current ($\times 10^4$ m^3 s^{-1}) (positive into the North Sea)
τ_w = Wind stress (N m^{-2})
Θ = Direction from which wind stress directed
Θ_{max} = Direction from which wind stress results in maximum transport towards the North Sea
= 202° (Pingree and Griffiths, 1980)
= 110–210° (Turrell *et al.*, 1990, 200° used in calculations)

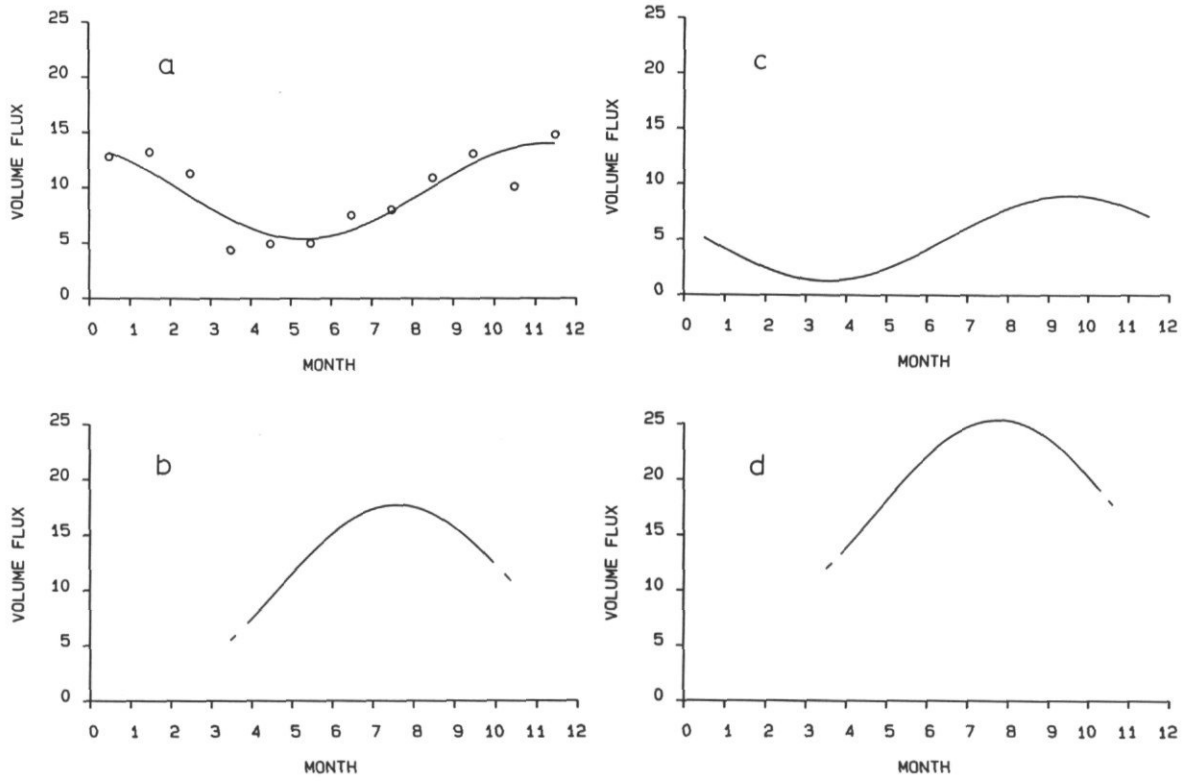

Figure 10. Annual cycles of volume transport derived from simple models of wind and density-driven currents. (a) Wind-driven transport in the Fair Isle current west of Orkney before entering the North Sea; (b) density-driven transport in the East Shetland Atlantic Inflow; (c) wind-driven transport in the East Shetland Atlantic Inflow; and (d) density-driven transport in the Fair Isle current in the northern North Sea. (Note: time is month − 1).

c = Non-wind-driven component ($\times 10^4$ m^3 s^{-1})
= −0.4 (Pingree and Griffiths, 1980)
= 0.4 (Turrell *et al.*, 1990)

m = Linear regression coefficient
= 33.0 (Pingree and Griffiths, 1980)
= 22.2 (Turrell *et al.*, 1990)

To examine the seasonal variation of the Fair Isle inflow, wind data have been obtained from Kirkwall, Orkney. Wind stress has been computed from wind speed and direction recorded at 0900 GMT each day over the period 1970–1990, and daily values of volume transport have been calculated. These values have then been averaged over each month of the year to derive monthly averages for the 20-year period.

The resulting mean annual cycle of volume transport into the North Sea via the wind-driven Fair Isle current is shown in Figure 10a. Each month inflow occurs, on average, owing to the predominance of southwesterly winds, while the annual cycle, exhibiting maximum inflow in December ($\sim 15 \times 10^4$ m^3 s^{-1}) and minimum in June ($\sim 5 \times 10^4$ m^3 s^{-1}), is the result of seasonal variation in the wind stress.

East Shetland Atlantic Inflow – non-wind-driven

Although data obtained during ACE after the autumnal breakdown of vertical stratification were limited in extent, it would appear that during the winter months the inflow was reduced by approximately 50% and may no longer have been confined as a narrow current east of Shetland (Turrell *et al.*, 1992). It was clear, however, that a significant non-wind-driven component was present, at least during the summer months.

Turrell (1992) employed an analytical model proposed by Svendsen *et al.* (1991) to estimate the density-driven component of the East Shetland Atlantic Inflow. This model assumes that a well-mixed water column lies inshore from a stratified ocean, and estimates the geostrophically balanced flow occurring within the well-mixed waters above the intercept of the offshore pycnocline with a sloping sea bed (Fig. 11). These conditions are observed both east of Shetland across the Atlantic inflow, and to some degree across the Fair Isle current within the North Sea, during the summer.

The model is a modified form of the Margules relation for geostrophic two-layer flow. Assuming that

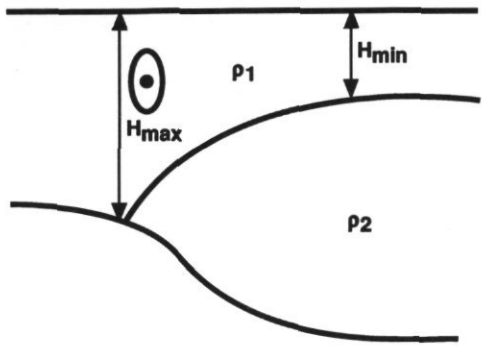

Figure 11. Schematic diagram showing the terms used in the two-layer analytical model of geostrophic transport (from Svendsen *et al.*, 1991).

$h_{max} \approx 100$ m, $h_{min} \approx 50$ m (Fig. 11) for both the East Shetland Atlantic Inflow and the Fair Isle current in the northern North Sea, then the equation reduces to the simple estimate of transport:

$$Q_{AW} = 35.8\ (\rho_2 - \rho_1) \times 10^4\ m^3\ s^{-1} \qquad (2)$$

where ρ_2 is the density of the offshore near-bottom water, and ρ_1 is the density of water within the vertically well-mixed current circulating cyclonically around the denser bottom water.

In order to use this model to estimate the possible seasonal variation of the magnitude in the Atlantic inflow, indices of northern North Sea densities were required. These follow those derived by Turrell (1992). Temperature and salinity observations from 1960 to 1990 from two stations inshore from the 200 m contour northwest of Shetland (Fig. 2) have been used to derive an index for the density of the inflowing Atlantic Water (AW(N) Index). Observations within a box in the centre of the northern North Sea (Fig. 2) have been used to derive an index for the density of the water lying offshore from the East Shetland Atlantic Inflow (the Central Northern North Sea Index – CNNS).

Monthly-mean densities have been computed over the period 1960 to 1990, and annual cycles of each parameter, A, have been found, using a least-squares fitting procedure, of the form:

$$A = A_{mean} + A_{amp} \sin\left(\frac{2\pi}{Y}(t - \Phi)\right) \qquad (3)$$

where A_{mean} = annual mean value; A_{amp} = amplitude of annual cycle; t = time in months (1 January = 0); Φ = phase shift (months); and Y = 12 (Maddock and Pingree, 1982).

The resulting annual density cycles are presented in Figure 12, while the coefficients describing the cycles are presented in Table 1.

The annual density cycles have been used in Equation (2) to calculate the seasonal variation of the density-driven component of the East Shetland Atlantic Inflow (Fig. 10b). The computed transport reaches a maximum in late August, with a magnitude of $\sim 18 \times 10^4\ m^3\ s^{-1}$. The assumptions made by the model obviously only apply during the months when offshore vertical stratification exists in the central northern North Sea, i.e. from April/May until October/November. Outwith this period the density-driven component will be that due solely to horizontal density gradients, and will be an order of magnitude smaller (Prandle, 1984).

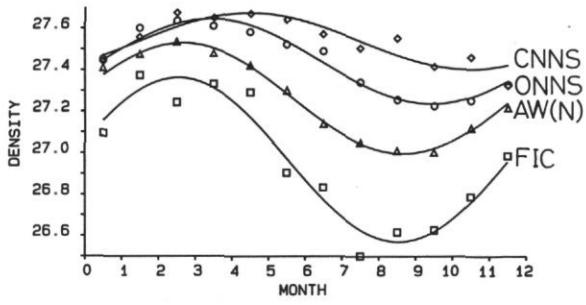

Figure 12. Annual cycles in northern North Sea densities. Density indices have been derived from: AW(N) Atlantic Water (North of Shetland) – two stations inshore from the 200 m contour north of Shetland; CNNS Central Northern North Sea water – box seen in Figure 2; ONNS Offshore Northern North Sea water – six easterly stations of JONSIS line (Fig. 2); and FIC Fair Isle current water – three westerly JONSIS stations. Full depth averages have been used for the AW(N) and FIC indices as the water is mixed on entry to the North Sea. The ONNS and CNNS are bottom water (>60 m) indices. (Note: time is month – 1).

Comparison of the inflow predicted by Equation (2) for the month of October ($\sim 14 \times 10^4\ m^3\ s^{-1}$) with the inflow deduced to have occurred east of Shetland in October 1987 (Fig. 8 – $\sim 35 \times 10^4\ m^3\ s^{-1}$) shows that only ~40% of the actual inflow may be explained by this simple model of the density-driven component. This discrepancy may be due partly to the smaller density difference between the inflowing Atlantic Water and the Central Northern North Sea water predicted by the mean annual cycles, compared to the difference that actually occurred in October 1987. For example, Turrell (1992) used the density difference observed in September 1990 to compute the density-driven component and produced a value of $\sim 21 \times 10^4\ m^3\ s^{-1}$.

There remains, however, a component of the inflow that is not density-driven. The regressions obtained by Pingree and Griffiths (1980) have been used to estimate the seasonal variation in the wind-driven inflow east of Shetland.

East Shetland Atlantic Inflow – wind-driven

Regression equations describing the wind-driven inflow to the North Sea were derived by Pingree and Griffiths

Table 1. Coefficients describing the annual cycles of density and transport within the northern North Sea. Units are density kg m^{-3} − 1,000, transport × 10^4 m^3 s^{-1}. Transports have been computed using simple models of density and wind-driven flow (see text). Negative values of transport indicate total outflow from the North Sea. The wind-driven component of the East Shetland Atlantic Inflow (ESAI) has been computed as being the difference between the total inflow between Shetland and Norway (SN) and the inflow within the Norwegian Trench (NT). Minimum values of density-driven transports are omitted as the two-layer model used to compute them is invalid during the winter months when no offshore vertical stratification exists.

	A_{mean}	A_{amp}	Phase lag Φ	Date of		Magnitude of	
				Max	Min	Min	Max
Density							
AW(N)	27.264	0.266	−0.33	20 Mar	20 Sep	26.998	27.530
CNNS	27.540	0.140	1.51	15 May	15 Nov	27.400	27.680
ONNS	27.443	0.204	0.44	14 Apr	14 Oct	27.239	27.647
FIC	26.967	0.395	−0.46	16 Mar	16 Sep	26.572	27.362
Transport – wind-driven							
FIC(WO)	9.70	4.29	−3.68	9 Dec	9 Jun	4.41	14.76
ESAI	5.13	3.83	−5.47	15 Oct	15 Apr	1.30	8.96
SN	−11.41	8.67	2.33	10 Jun	10 Dec	−20.08	−2.74
NT	−16.54	11.35	1.80	24 May	24 Nov	−27.89	−5.19
Transport – density-driven							
FIC(NS)	17.04	8.31	4.76	22 Aug	–	–	25.35
ESAI	9.88	7.82	4.62	19 Aug	–	–	17.70

(1980) for the total inflow between Shetland and Norway:

$$Q_{SN} = -45 \cos (\Theta - 199°) \quad (4)$$

and the total inflow within the Norwegian Trench:

$$Q_{NT} = -58 \cos (\Theta - 228°) \quad (5)$$

where Θ is the direction from which wind stress is directed.

The 20-year record of wind at Kirkwall has been used to derive annual cycles of wind-driven inflow using these two regression equations (Table 1). The difference between the two cycles provides an approximation of the inflow that occurs between Shetland and the western edge of the Norwegian Trench (Fig. 10c). The resulting cycle describing the inflow of Atlantic water east of Shetland demonstrates a maximum influx for winds from 280°. The seasonal cycle of transport reaches a minimum in April (~1 × 10^4 m^3 s^{-1}) and a maximum in October (~9 × 10^4 m^3 s^{-1}).

The total inflow east of Shetland (wind- plus density-driven) predicted for October, when observations are available for comparison (Fig. 8), is ~23 × 10^4 m^3 s^{-1}. This is approximately 66% of that observed, and does not include the additional inflow of ~21 × 10^4 m^3 s^{-1} that the observations suggested in the central areas of the northern entrance to the North Sea, offshore from the actual East Shetland Inflow.

Thus the actual magnitudes predicted by the models may be in error when compared to any particular year. In the discussion below, however, the models are used to indicate the form of the seasonal variability in the inflows in order to summarize the new concepts of northern North Sea circulation that have arisen from the results of ACE and subsequent studies east of Shetland.

Fair Isle current within the North Sea

Turrell *et al.* (1990) found that within the northern North Sea during the winter months, the Fair Isle current behaves similarly to the current as observed to the west of Orkney, before it enters the North Sea. Thus a linear response to wind stress was found, as described by Equation (1).

During the summer months, however, a significant non-wind-driven component became apparent. The model of Svendsen *et al.* (1991) may be applied to that part of the Fair Isle current flowing south from the Orkney/Shetland passage close to the 100 m contour. Hydrographic observations along the EC line during ACE (Fig. 5) demonstrate that the assumptions of the model were valid, with a less-dense, vertically well-mixed current lying inshore from and above a dense bottom layer.

Density indices of the water types involved have been derived using data obtained along the JONSIS line, east of Orkney, over the period 1970–1990 (Fig. 2). An index of the water within the Fair Isle current (FIC Index) has been obtained from the three most westerly stations, while the density of the offshore bottom water (ONNS Index) has been characterized by averages of values from the six most easterly stations (Turrell, 1992).

When the indices are applied in Equation (2) the resulting seasonal development of the density-driven Fair Isle current indicates a maximum southerly transport of ~25 × 10^4 m^3 s^{-1}, occurring in late August. The

predicted magnitude for October ($\sim 22 \times 10^4$ m^3 s^{-1}) agrees well with the transport deduced from the observations obtained during ACE (Fig. 8).

Conceptual annual cycle of northern North Sea circulation

The results from ACE, in conjunction with the results of the simple models of wind and density-driven transports presented above, allow a new conceptual model of the annual circulation cycle of the northern North Sea to be proposed.

February–April (winter)

At this time, since vertical stratification is absent in the northern North Sea, only weak horizontal density gradients are present. Density forcing will therefore be at a minimum. However, the densities of the different water types are all at or approaching their maximum values, with the CNNS and ONNS water types being most dense, FIC water least dense, and Atlantic water at an intermediate density. The difference between the density of the Atlantic water found north of Shetland and the water lying within the entrance to the northern North Sea (CNNS water) is at a minimum, and thus inflows driven by the wind or external oceanic forcing may occur most readily.

Within the EC box all water types are at a similar temperature of ~6–7°C (Fig. 13), while the temperature of the Atlantic water north of Shetland remains above ~8°C (Turrell, 1992). This implies that little direct oceanic inflow is occurring. Water reaching the EC line is either the cool, mixed Atlantic and coastal waters of the Fair Isle current, or water from the northern North Sea which has been present throughout the winter and has therefore cooled significantly.

The wind-driven inflow of the Fair Isle current is reducing during this period from its maximum in December towards its minimum in June. The simple model of wind-driven inflow indicates that the Fair Isle current transport is $\sim 7 \times 10^4$ m^3 s^{-1} during this period, that is, approximately ~50% of its December value. Transport within the Fair Isle current in the North Sea is wholly wind-driven, and of a similar magnitude to that entering between Shetland and Orkney, although Turrell *et al.* (1990) noted a small increase in transport within the current in the North Sea compared to that west of Orkney ($m = 35$ compared to $m = 22$ in Equation (1)). The additional water will have been entrained into the current during its passage between Orkney and Shetland, and south from there.

East of Shetland weak horizontal density gradients exist, and any inflow occurring will be broad and diffuse. The simple estimate of wind-driven inflow suggests a very small value of $\sim 1 \times 10^4$ m^3 s^{-1}, with wind-driven transport east of Shetland reaching a minimum in April.

May–July (spring)

As solar heating increases and wind stress declines, vertical stratification is established throughout much of the northern North Sea, isolating the cool dense bottom waters from the warming effects of solar heating. This thermal stratification permits coastal water to spread west from the coast of Norway and the Skagerrak, reducing the surface salinity over much of the north-eastern North Sea and hence enhancing vertical density gradients and further reducing vertical mixing across the pycnocline.

The Fair Isle current itself, however, remains vertically well-mixed owing to intense tidal mixing as it passes between Orkney and Shetland. West of Orkney, the current remains wind-driven throughout the year, reaching a minimum during this period ($\sim 5 \times 10^4$ m^3 s^{-1}). Salinity observations made during ACE suggest that the current at least doubles in volume flux as it enters the North Sea in the summer months, through the Orkney/Shetland gap, as a result of less saline inshore waters being entrained.

Thus the Fair Isle current enters the North Sea as a well-mixed flow of reduced density lying inshore from a stratified sea with a dense bottom layer. The result is a predominantly density-driven flow within the North Sea, dynamically detached from the current west of Orkney. Variability in the wind-driven inflow and in the local wind forcing modifies the current in a complex way, altering both the supply of water from west of Orkney and the surrounding three-dimensional density structure. Volume enhancement of the density-driven current may take place with the entrainment of coastal water and also water from the northern part of the North Sea.

While predominantly wind- and density-driven cyclonic circulation exists around the basin of the northern North Sea on a time scale of weeks the circulation can reduce and even reverse under the influence of internally and externally generated storm surges on a time scale of days (Backhaus, 1984; Turrell and Henderson, 1991). These events will result in increased horizontal mixing and in a reduction in the density-driven current.

A similar situation to that of the Fair Isle current in the North Sea exists at this time east of Shetland, with offshore vertical stratification developing. North of Shetland another area of intense tidal mixing exists (Pingree and Griffiths, 1978). Wind and externally driven inflow to the North Sea entering east of Shetland will thus be vertically well-mixed and the dynamics described by Svendsen *et al.* (1991) will again be applicable. The resulting density-driven East Shetland Atlantic Inflow will inject warm, saline water into the northern North Sea, circulating around the cold, dense offshore bottom water. This feature can be seen in the observations made within the EC box at this time, as the inflowing Atlantic water is much warmer than the still cool coastal waters (Fig. 13).

August–October (summer)

During the summer months, the vertical stratification in most areas of the northern North Sea is enhanced. Sporadic wind events may increase the mixed layer depth while strengthening gradients at the pycnocline. Density differences between the waters within the Fair Isle current and the East Shetland Atlantic Inflow, and those offshore from them increase to maximum values in the late summer. Thus the density-driven fluxes within the two currents as predicted by the simple analytical model, reach their maximum values during this period. The wind-driven components of the inflows are increasing towards their autumnal maxima.

A fully developed T-S profile in the EC box (Fig. 13) shows warm fresh coastal water lying inshore from the mixed Fair Isle current water entering the North Sea through the Orkney/Shetland passage. Water of recent Atlantic origin brought in as a result of the East Shetland Atlantic Inflow lies offshore from this current and results in the salinity maximum in the T-S diagram. Within the bottom waters of the central northern North Sea lies the Cooled Atlantic Water, which retains the temperature and salinity characteristics observed in March/April. During these months the presence of these dense bottom waters is essential, driving the cyclonic circulation in the less-dense vertically well-mixed waters which surround them.

November–January (autumn)

During autumn the wind stress increases and results in increased vertical mixing in the central areas of the northern North Sea, ultimately removing vertical stratification. As the vertical structure is eroded, the density-

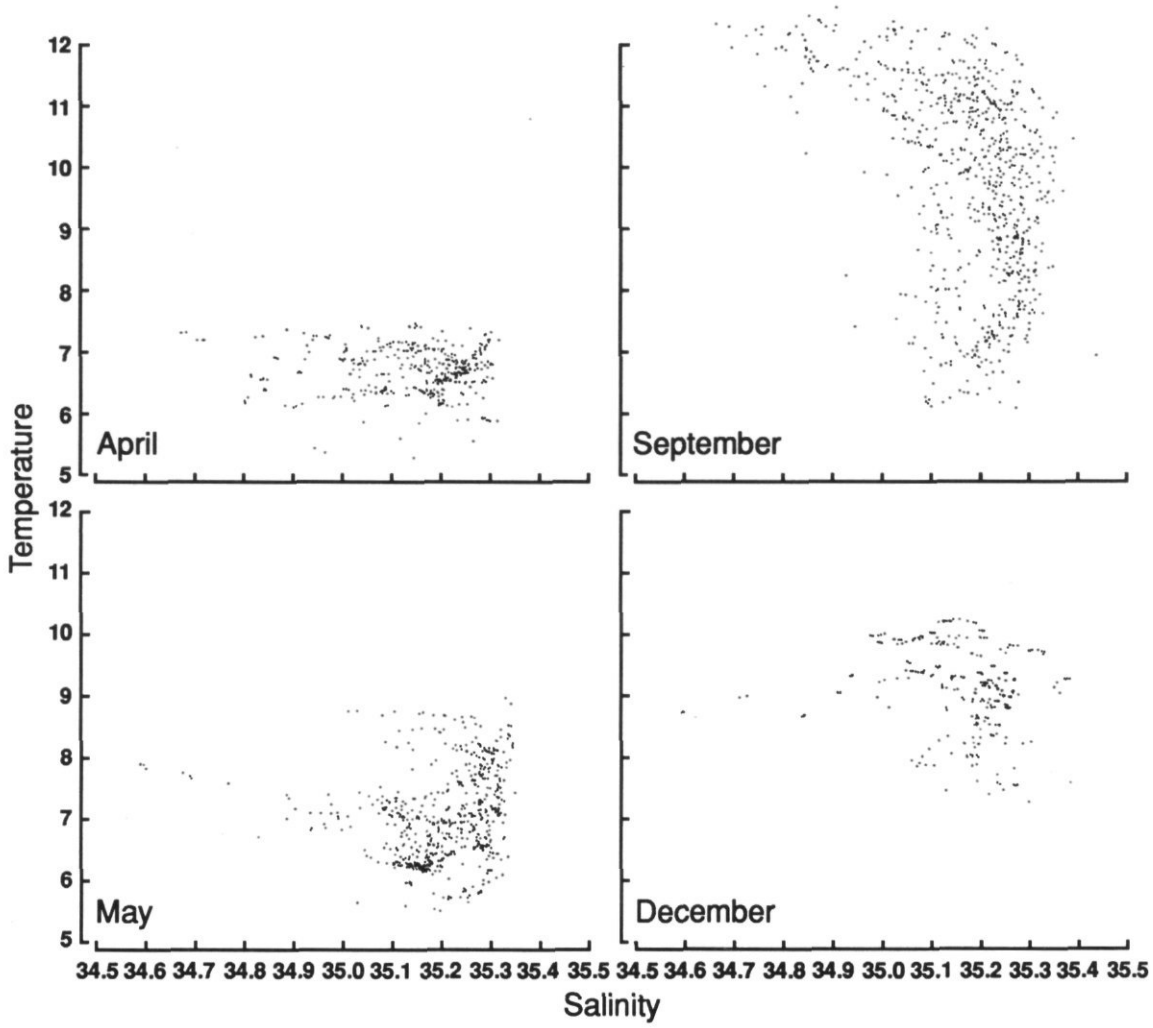

Figure 13. Monthly temperature–salinity diagrams constructed from observations made within the East Coast (EC) box (Fig. 2) throughout the period 1960–1990.

driven transports within the northern North Sea cease. The increasing, predominantly westerly winds, however, increase the wind-driven transports. Typical values indicated by the model of wind-driven flow are $\sim 14 \times 10^4\ m^3\ s^{-1}$ for the Fair Isle inflow and $\sim 9 \times 10^4\ m^3\ s^{-1}$ for the inflow east of Shetland.

While the surface temperatures decrease during autumn, the temperatures of the bottom waters in the central northern North Sea increase as warmer water is mixed down from the surface. This can be seen in the observations from the EC box in December (Fig. 13). The characteristic temperature and salinity of the central lens of dense bottom CAW water can no longer be seen as temperatures across the area equilibrate. During the remainder of the autumn and during the winter the temperatures slowly decline in the nothern North Sea to their minimum values prior to the onset of the next annual cycle. Cessation of the East Shetland Atlantic Inflow during the autumn is demonstrated by the disappearance of saline Atlantic water which maintains higher temperatures throughout the winter.

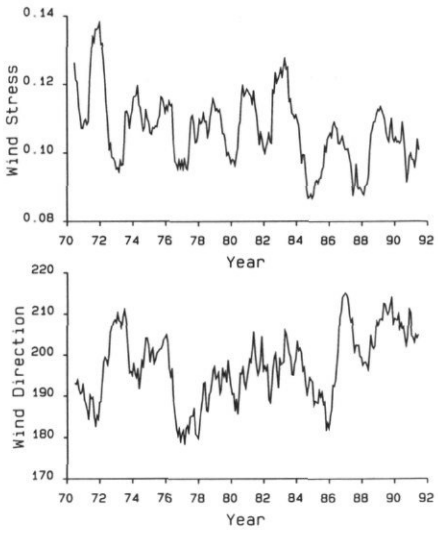

Figure 14. Twenty-year record of wind stress (Nm^{-2}) and direction (from) for wind recorded at Kirkwall on Orkney. A one-year running mean has been applied to the data to reduce seasonal noise.

Interannual variability

The interannual variability of the wind-driven components of the two western inflows to the northern North Sea may be examined using the 20-year record of wind data from Kirkwall, in conjunction with the simple directional-response transport models. The time series of monthly-mean wind stress may be seen in Figure 14, along with the mean directions from which the stress occurs. A one-year running mean has been applied to the data in order to reduce the seasonal signal (Furnes, 1992).

Long-term (>2 years) trends were apparent in the wind forcing, particularly in the evolution of the mean direction of the wind stress. During 1970 and 1971, the wind stress gradually backed, almost reaching a southerly direction, with variable strength. This was followed by a severe veering of the wind, returning to a southwesterly direction and with greatly increased magnitude during 1972. There then followed a three-year period of again gradual backing of the wind until 1976 when the wind stress reduced and rapidly backed towards south. From 1977 until 1982 the mean direction slowly veered again from south towards southwest, with the magnitude of the stress reaching a maximum in the winter of 1982/1983. During the remainder of 1983 the wind again backed with decreasing strength until the beginning of 1986, during which a marked veering occurred. The wind direction reached its most westerly during 1987 and then backed until mid-1988 when it again veered towards the west until the end of 1991. Wind stress increased from the end of 1987 until 1989 when it began to decrease.

The changes in wind stress magnitude and direction combine to produce the interannual variability of the wind-driven component of the Fair Isle Current (Fig. 15a) and the East Shetland Atlantic Inflow (Fig. 15b). The Fair Isle Current demonstrated a sustained maximum from late 1971 until 1973. This was initially due to high wind stress, but was maintained by the veering of the wind from south back towards the optimum direction for the Fair Isle inflow of 200–210°. The minimum transport that followed in 1976/1977 was due to both reducing wind stress and backing of the wind to the south. Increase in the transport in 1979/1980, followed by a rapid decline was associated with changes in wind direction rather than strength, while the subsequent maximum transport in 1982/1983 was associated with both high wind stress and a direction from the optimum southwest. Declining wind stress resulted in a reduced inflow in 1985, followed by a period of increased transport into the North Sea in 1986 caused by changes in both wind strength and direction. During the end of the decade mean annual transport increased owing to the veering of the mean wind direction towards the southwest and the simultaneous increase in wind stress.

Similar trends over the 20-year period are seen in the wind-driven component of the East Shetland Atlantic Inflow (Fig. 15b). The mean wind direction at Kirkwall is, however, almost in quadrature with the predicted optimum wind direction which results in the maximum inflow within the East Shetland current (280°), and hence much greater variability of the transport results. Backing of the wind beyond 190° results in reversed flow east of Shetland.

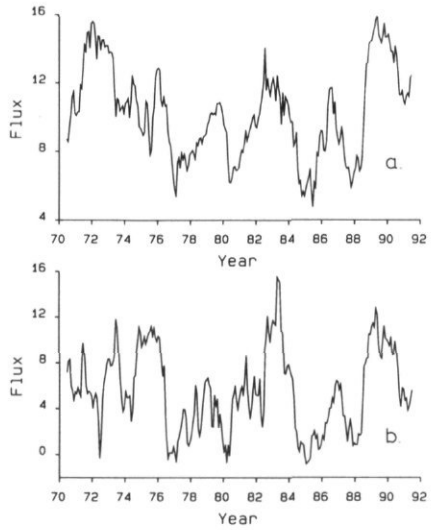

Figure 15. The wind-driven component of the: (a) Fair Isle Current; and (b) East Shetland Atlantic Inflow as predicted by the simple diagnostic models of transport derived from Pingree and Griffiths (1980). Units are $10^4\ m^3\ s^{-1}$.

Discussion

The results presented here, obtained during the Autumn Circulation Experiment and from subsequent experiments in the northern North Sea, permit a conceptual annual circulation cycle of the northern North Sea to be proposed. A predominantly wind-driven inflow towards the North Sea, directed through the Orkney/Shetland passage, occurs all year round, reaching a maximum in December and a minimum in June. The current flowing out of the Fair Isle area during the winter is also wind-driven, and circulates around the basin of the northern North Sea cyclonically. In the winter, a weak wind-driven inflow occurs from the north, to the east of the Shetland Islands.

The onset of seasonal stratification during the spring brings important changes to the dynamics of the northern North Sea. Offshore waters become vertically stratified, with cool dense bottom waters, isolated from direct erosion by wind mixing, lying beneath warm and increasingly less saline surface waters. The inflowing water remains vertically well-mixed as areas of intense tidal mixing lie at the points of entry of the inflows, both between Orkney and Shetland and north of Shetland (Pingree and Griffiths, 1978). Baroclinic forcing becomes increasingly important within the basin of the North Sea as the mixed layer depth increases, and density differences between the inflowing and resident waters become enhanced. The wind-driven inflow west of Orkney becomes insufficient to balance the transport within the Fair Isle current flowing south from the Orkney/Shetland passage within the northern North Sea, and additional volume flux is supplied by inshore coastal waters entrained into the current.

The next transitional period is the autumn. As the wind stress increases and solar heating declines vertical mixing results in an increase in the mixed layer depth, with an associated removal of the deep water that has survived the predominant horizontal erosion throughout the summer in the central northern North Sea. The removal of the offshore two-layer structure results in the reduction of the baroclinic forcing and the circulation reverts to a predominantly wind-driven one.

This annual circulation cycle of the northern North Sea differs from that proposed by Steele (1957). He described a large ingress of dense Atlantic water to the North Sea from the Faroe–Shetland Channel during the autumn, mixing with old water remaining in the northern North Sea. Towards March winter cooling of the water within the North Sea increased its density to a point where it halted further Atlantic inflow. From the present study it is clear that water at the entrance to the North Sea is less dense than northern North Sea water all year round, with only the density difference being least in March (Fig. 12). The autumnal ingress of Atlantic water is in fact driven by the density difference with the inflowing water becoming vertically mixed at its entrance to the North Sea resulting in the buoyancy driven flow around the bottom layer "lens" of cold, dense old northern North Sea water (CAW).

The biological consequences of this circulation cycle are profound, and affect all trophic levels within the northern North Sea ecosystem. While the offshore seasonal stratification results in nutrient depletion in the surface waters of the central northern North Sea following the spring bloom, the vertically well-mixed inflows result in a continuous supply of nutrients which maintains productivity at current boundaries throughout the summer. Thus both primary and secondary production is enhanced in the northern North Sea (Dooley, 1981).

The circulation itself directly affects certain fish species through the advection of their planktonic stages. For example the transport and dispersal of herring larvae, which hatch from spawning grounds to the north and west of Scotland in the autumn (Heath, 1989), are dependent upon the circulation existing at the time. It has been suggested that, in order for them to survive to join the adult stock, herring larvae must cross the northern North Sea within the transverse currents to arrive at their nursery grounds in the Skagerrak and southern North Sea (Bartsch *et al.*, 1989). Failure to do so may be brought about by interannual variability in the circulation, and this mechanism has been suggested as being partly responsible for the failure of the herring stocks in the late 1970s (Corten, 1990; Turrell, 1992).

Similarly, Turrell (1992) has proposed that spring-spawned sandeels to the east of Shetland may become lost to the local stock if they are entrained into the East

Shetland Atlantic Inflow. Long-term changes in the Shetland sandeel stock, and hence the seabirds which feed upon them, may have been caused by the annually varying strength of the circulation. It is particularly interesting that this stock demonstrated a dramatic increase during the same period that herring declined in the North Sea.

The mechanisms controlling the year to year variation of Atlantic inflow to the North Sea remain unknown. It has been suggested that hydrographic changes which occurred in the mid to late 1970s associated with what is now known as the Great Salinity Anomaly (Dickson *et al.*, 1988), coupled with a weak Gulf Stream transport and reduced westerly winds over the North Sea, may have been responsible for reduced Atlantic inflows, and that this may account for the biological changes observed (Turrell, 1992). In the absence of any long-term monitoring of the circulation of the North Sea, interannual variability suggested by the biological evidence must be addressed using models of the circulation driven by climatic records of weather and hydrographic parameters. From the simple models employed here it is evident that there are significant long-term trends in both wind stress magnitude and direction which affect the inflows to the North Sea, and hence its temperature, salinity, and flushing characteristics (Furnes, 1992).

The critical planktonic phases of both herring and sandeels occur at times when the circulation of the northern North Sea is undergoing change. Thus the timing and magnitude of specific events within an individual year may alter biological development within the northern North Sea. For example, the timing of the onset of seasonal stratification, which enhances the circulation through baroclinic forcing, in relation to the spawning time of Shetland sandeels may be crucial in determining the success of the stock. The present day numerical models which include baroclinic forcing generally employ mean density fields derived from average values of observations obtained over many years (Backhaus, 1984). In order to simulate correctly the annual circulation cycle of the northern North Sea, and its effects upon primary, secondary, and higher levels of production, it will be necessary to develop density-evolving models driven by real weather.

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Climatic variability in the North Sea

Einar Svendsen and Anne Karin Magnusson

Svendsen, E., and Magnusson, A. K. 1992. Climatic variability in the North Sea. – ICES mar. Sci. Symp., 195: 144–158.

Twenty-two years of hydrographic data taken during the summer over most of the northern and central North Sea show large year to year variations in the distribution of high salinity Atlantic Water, heat, and a stable low salinity surface layer. From these data we have extracted well defined integrated parameters claimed to indirectly describe/quantify the variability of: (1) the Atlantic Water inflow, (2) the heat content as a combined effect of this inflow and the surface heat flux, and (3) the area strongly affected by the surface spread of Norwegian Coastal Water. Some of these parameters are found to be strongly affected by local weather. With the use of multiple regression analysis it has been possible to quantify connections and dynamical responses between these ocean variables and derived meteorological variables such as wind stress, turbulence effect, and cloud coverage. The extracted ocean and meteorological variables are assumed to be of prime importance for the biological productivity of the North Sea. For example, it is found that the recruitment success of some of the fish stocks seems to be quantifiable based on these climatic variables prior to and during the larval stages. These results are interesting with respect to possible better management of our fishery resources.

Einar Svendsen: Institute of Marine Research, PO Box 1870 N–5024 Bergen, Norway; Anne Karin Magnusson: The Norwegian Meteorological Institute, Allegaten 70, N–5007 Bergen, Norway.

Introduction

Most marine biologists agree that the physical climate is important for the ocean ecology, but relatively few quantitative interrelations are found or published. This is because many of these investigations consider only one single physical parameter at a time (very often temperature), which in addition is not necessarily a prime ecological steering factor. Corten (1990) suggests that most of the observed changes in fish stocks could be explained by a theory assuming a long-term reduction of the inflow of Atlantic Water (AW) to the North Sea (NS) during the period 1960–1980, and an increase of this inflow in later years. However, he also states that there is as yet no physical evidence that such a change occurred.

We believe that one of our derived (from hydrography) parameters indeed represents an index of the year-to-year variability of the AW inflow. Kamp (1990) also created a year-to-year index of the inflow of AW to the North Sea using a different method and different hydrographical data, and there are clear similarities in the results.

The Institute of Marine Research (IMR) in Bergen has yearly expeditions around June/July in the northern and central North Sea. About 300–500 hydrographic (CTD) stations are taken in this period, giving in general a good areal coverage at least north of 57°N. These salinity and temperature data for the years 1968 to 1990 are the physical basis for this study, together with wind and cloud observations (obtained from the Norwegian Meteorological Institute, DNMI) at Utsira, an island off western Norway.

One of our derived parameters is associated with the heat exchange, which Colebrook and Taylor (1984) indicate as being related to a 3–4-year periodic fluctuation in the plankton abundance. Colebrook (1985) also indicates that the influence of wind and temperature on zooplankton abundance might primarily be through processes involved in vertical mixing and stability, both represented by our derived parameters.

An overall goal is to quantify the importance of the North Sea climate variability on biological variability, especially the recruitment of some North Sea fish stocks. To find such relations we have derived some physical parameters (indirectly describing volume and heat fluxes, local wind impact, etc.) believed to be the most important for primary and secondary production and the early stages of fish.

In this paper we aim to describe the local North Sea

climate variabilities through these parameters derived from observations in the North Sea and locally measured weather, and to quantify possible interrelations to explain some of the ocean climate and biological variabilities. The main part of the fish-climate results are separated into another paper (Svendsen *et al.*, 1991).

Primary physical variables

The AW inflow is not just important with respect to heat. It also brings large amounts of nutrients and zooplankton which are of great importance for the high productivity of this ocean area. In this respect the local wind climate will influence the amount of nutrients available for primary production due to turbulent mixing, upwelling etc., in addition to steering much of the near surface circulation (Sætre *et al.*, 1988), residual times, and contact rates between for instance prey and predators (Sundby and Fossum, 1990).

In addition to the variability of the inflowing AW and the wind conditions, the variable winter surface heat flux (cooling) has an impact on the subsurface water heat content the following year, being important for instance for the bottom water exchange in the Skagerrak (Ljøen, 1981). Also the stability of the water has been reported to be important, especially for the primary and secondary production (Dickson *et al.*, 1991; Melle and Skjoldal, 1989). The last, but not least important factor for some biological processes is the variable light conditions especially during spring bloom (Sakshaug and Skjoldal, 1989) and also important for the prey/predator interactions, depending on visibility (Aksnes and Giske, 1990). The light conditions are especially sensitive to the concentration of low and medium clouds, which also to a large extent regulates both cooling during winter and heating during summer.

In short, we have been searching for oceanographic parameters quantifying the year-to-year variability of the:

(1) Inflow of AW (both subsurface, AW_{deep}, and near-surface, AW_{surf}).
(2) Effect of winter cooling on subsurface water temperature/heat content (AW_{heat}).
(3) Near surface stability.

Additional independent variables being used to explain some of the above variabilities and observed biological variabilities are the:

(4) Wind (monthly/seasonal mean) stress components, STRESSN(orth), STRESSE(ast) and the turbulent wind effect, EFF(cubed wind speed).
(5) Cloudiness (monthly mean of low and medium, CLM).

Neither the near surface sea temperature nor the air temperature has been considered since we do not have mean values representative for the year-to-year variability of the northern North Sea. (Clearly, there are connections between these temperatures and wind speed/direction and cloudiness, so the effect is not totally left out.) Another parameter being left out is the supply of fresh water (with respect to stability), either through direct rainfall or through river runoff. Such data will be considered in future work.

Neither direct measurements nor realistic model results of the inflow of AW to the North Sea exist, and the directly measured salinity and temperature are sensitive to short time variabilities and to the characteristics of the inflowing AW. We believe that the volume (areal coverage over a certain depth range) of AW measured around June/July is a good index for the integrated flux during the previous months/half a year. Thus the fraction of the total area north of 57°N covered by high salinity water (greater than specified salinities) averaged from 50 to 200 m depth (or bottom) (AW_{deep}) are used as indirect measures of the influx of AW to the North Sea; 50–200 m depth is chosen since this represents the depth range of the core of the main flux along the Norwegian Trench (NT), and is relatively undisturbed by the fresher Norwegian Coastal Water (NCW) at the surface.

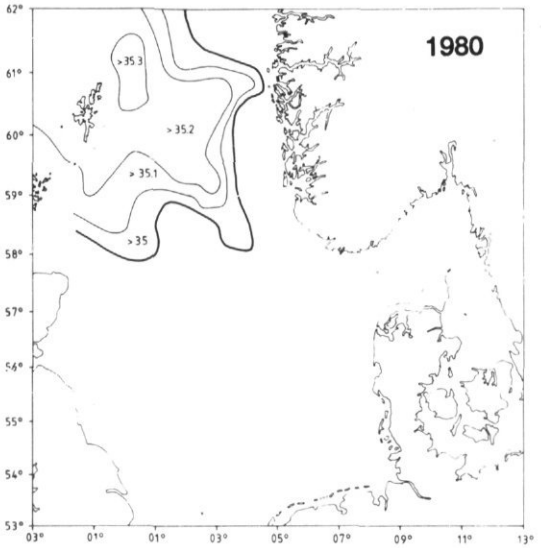
However, later research (Turrell *et al.*, 1991; Svendsen *et al.*, 1991) has revealed a significant inflow of AW also near the surface just east of Shetland, and a similar areal coverage has been estimated on averaged data from the upper 30 m (AW_{surf}).

Similar estimates on the temperature (greater than specified values) at the level of 50–200 m depth (AW_{heat}) give an indirect combined measure of the heat influx and the effect of cooling during the previous winter. No information has been used from near surface temperature measurements since these are very sensitive to the observation time in summer.

The stability is similarly indirectly parameterized by the areal fraction covered by NCW (NCW_{34}) estimated from the upper 30 m, average salinity being less than 34.0.

The above-mentioned yearly areal distributions are shown in Figs. 1, 2, and 3 for the 1980s, including 1990. During this period the minimum inflow occurred in 1980 connected with the ending of the "Late 1970s Anomaly" (Dickson *et al.*, 1988); the maximum inflow occurred in 1990 (Fig. 1). Clearly seen is the general subsurface transport of AW towards the Skagerrak along the NT (Fig. 1), and for most of the years the maximum southward displacement of AW in the western North Sea is restricted by the east to west Dooley Current at about 58°N (Svendsen *et al.*, 1991).

The temperature distributions (Fig. 2) also show large variations, and it is worth noting the drastic increase in temperature in recent years. From 1988 to 1990 the average temperature increased by nearly 1°C in the central parts of the area, clearly caused by the extremely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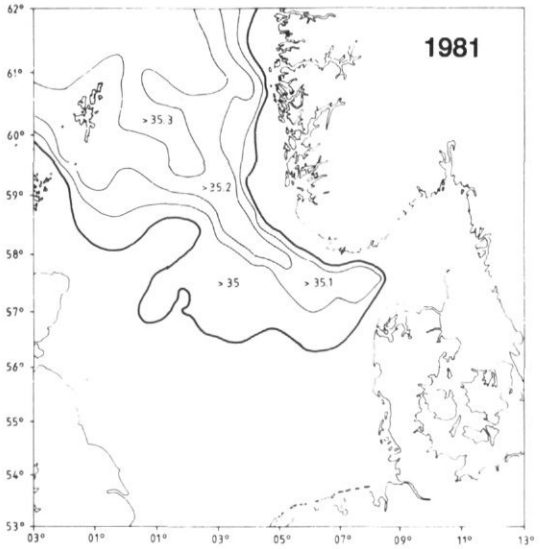
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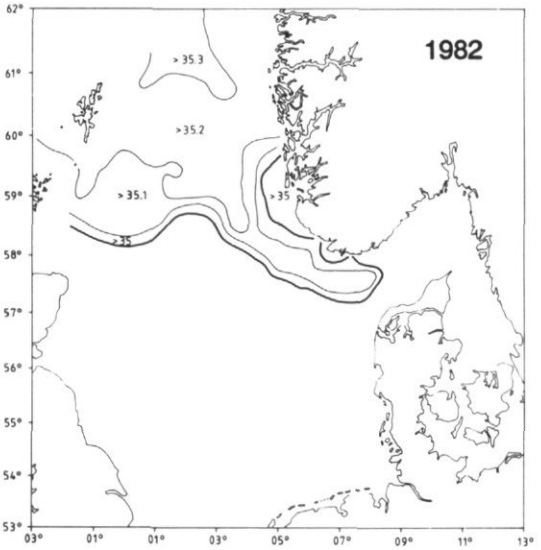
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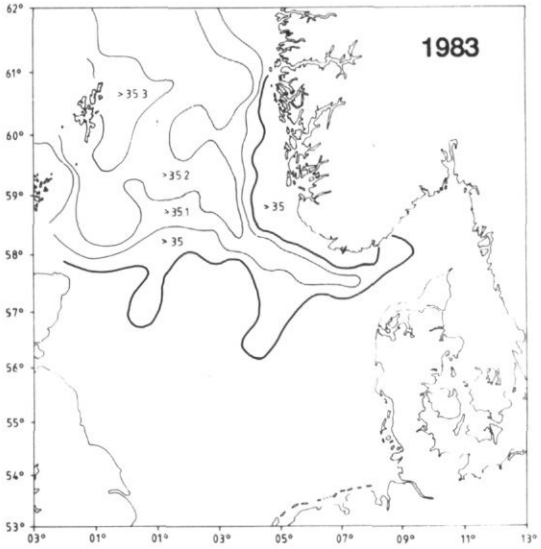
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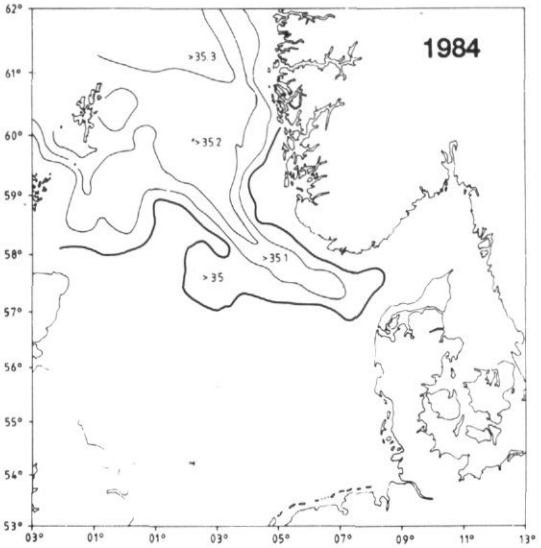
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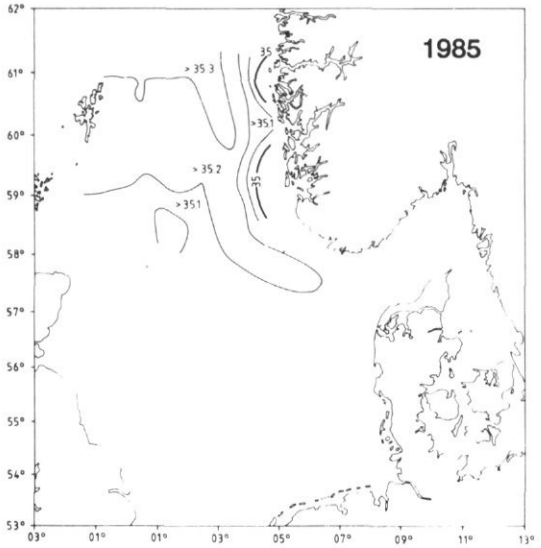
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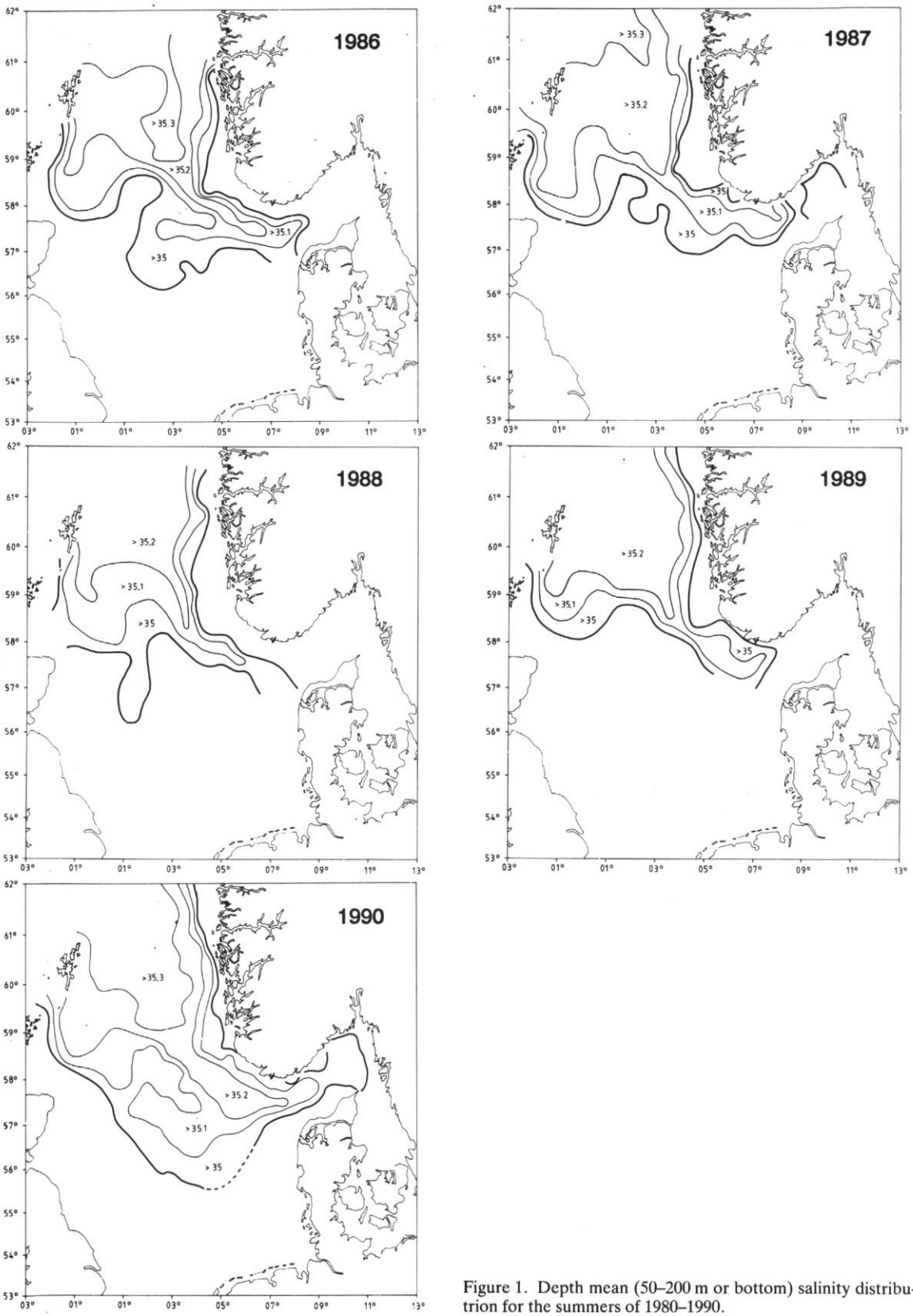
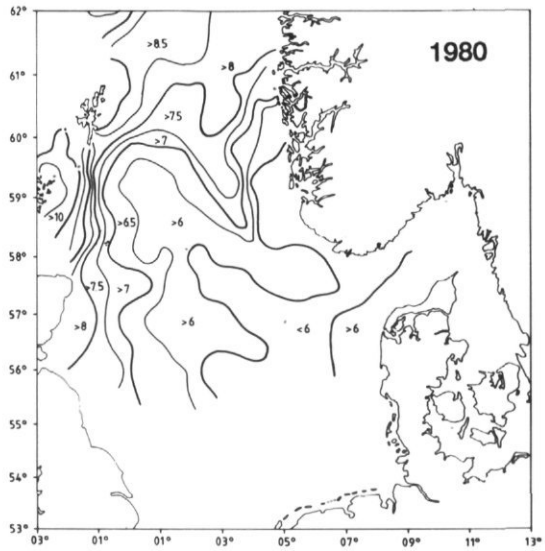
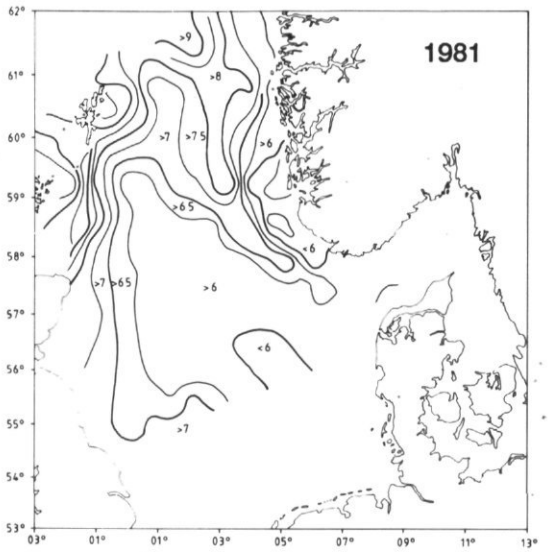


Figure 1. Depth mean (50–200 m or bottom) salinity distribution for the summers of 1980–1990.


1980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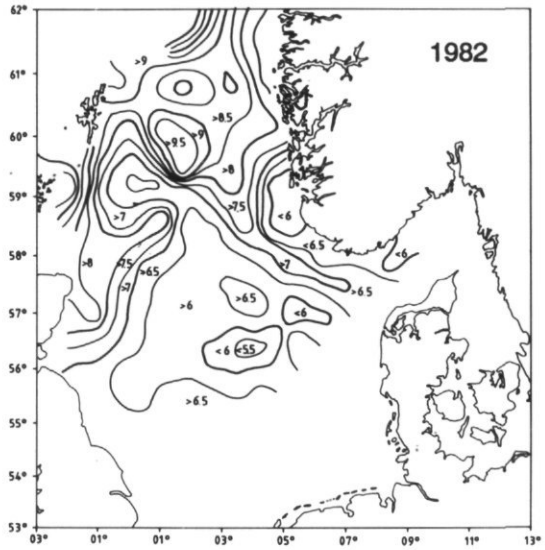
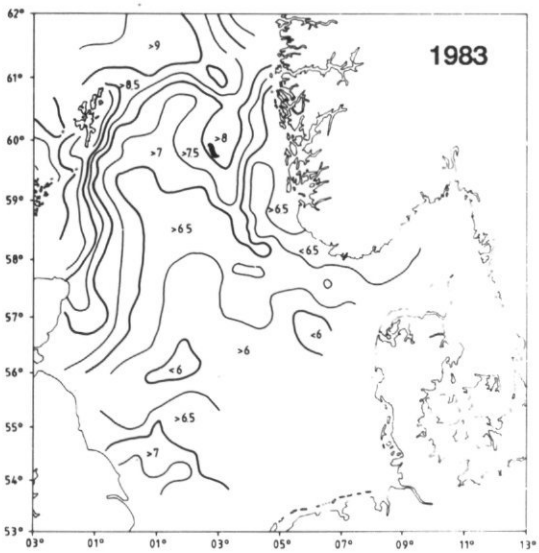
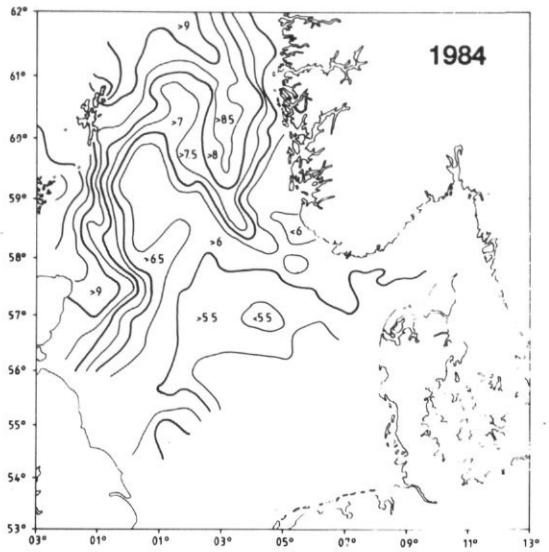

1982


1983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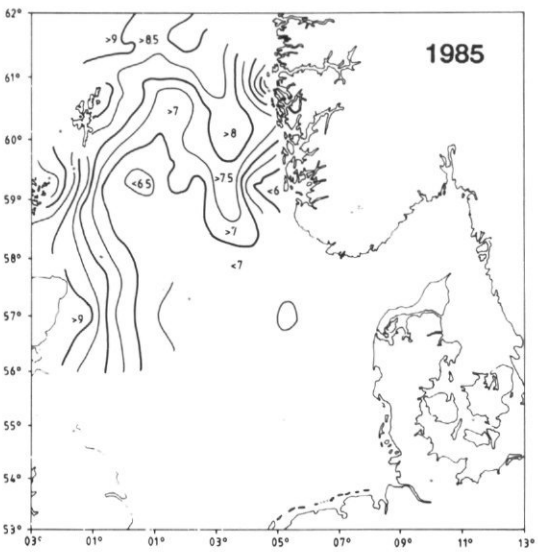

1985

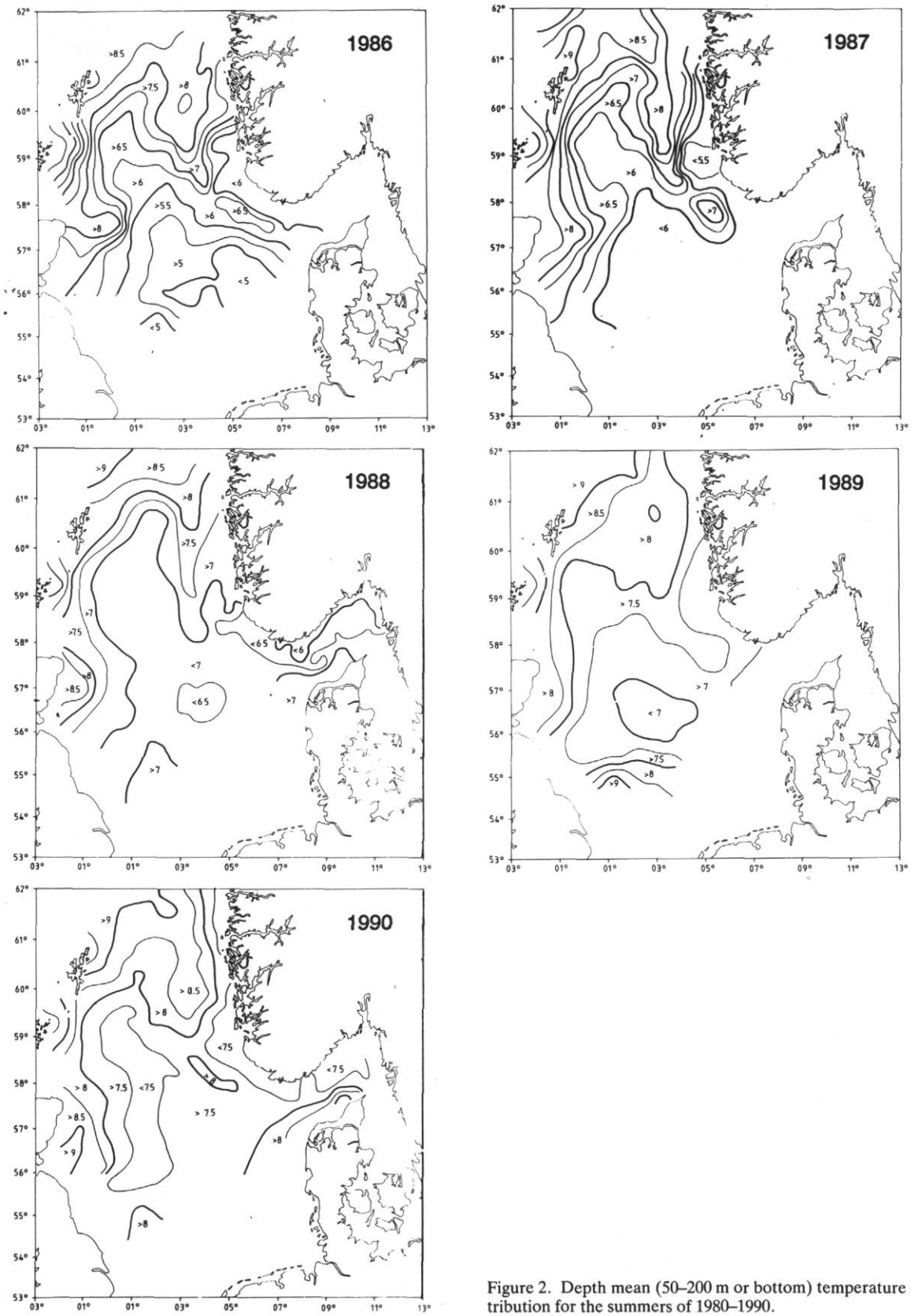
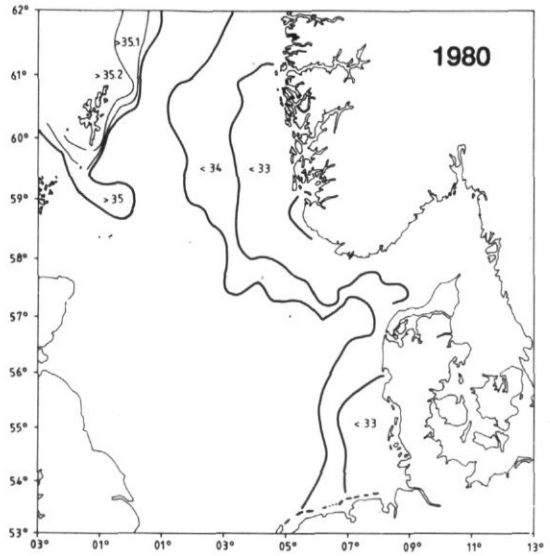


Figure 2. Depth mean (50–200 m or bottom) temperature distribution for the summers of 1980–1990.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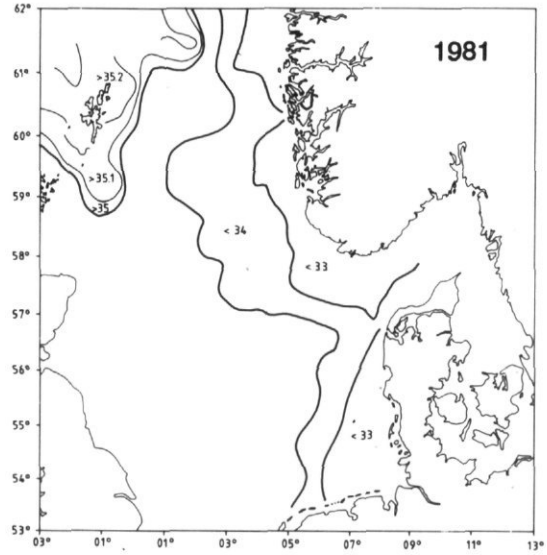
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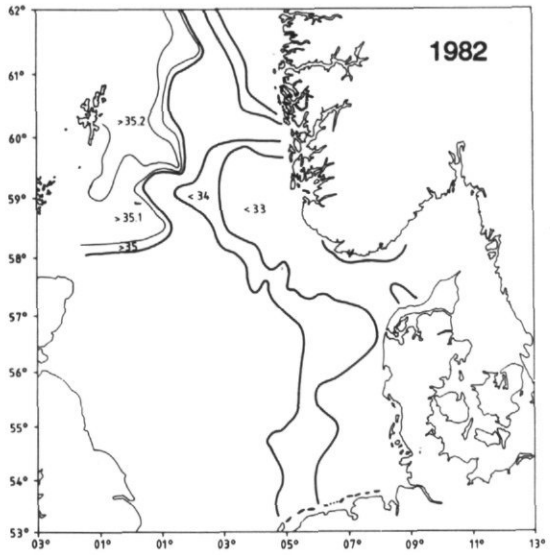
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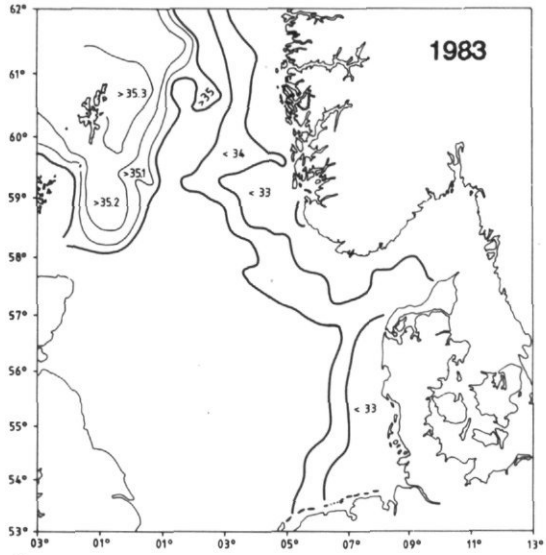
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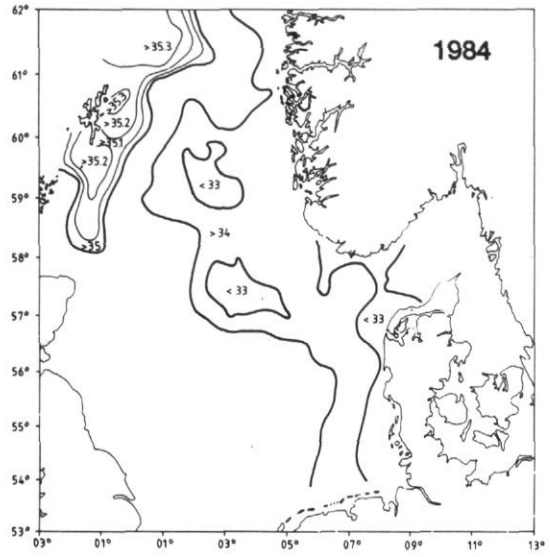
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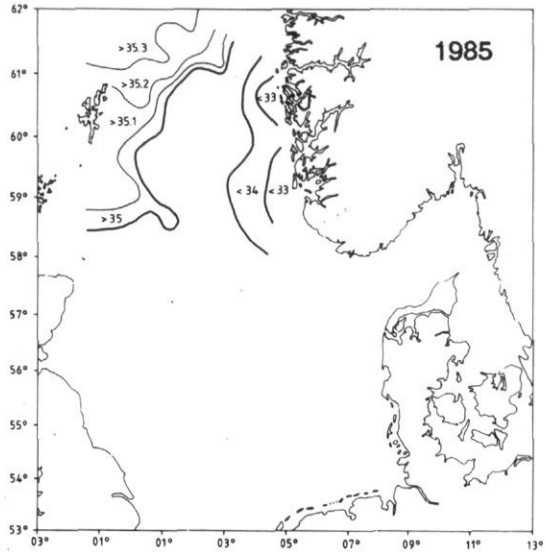
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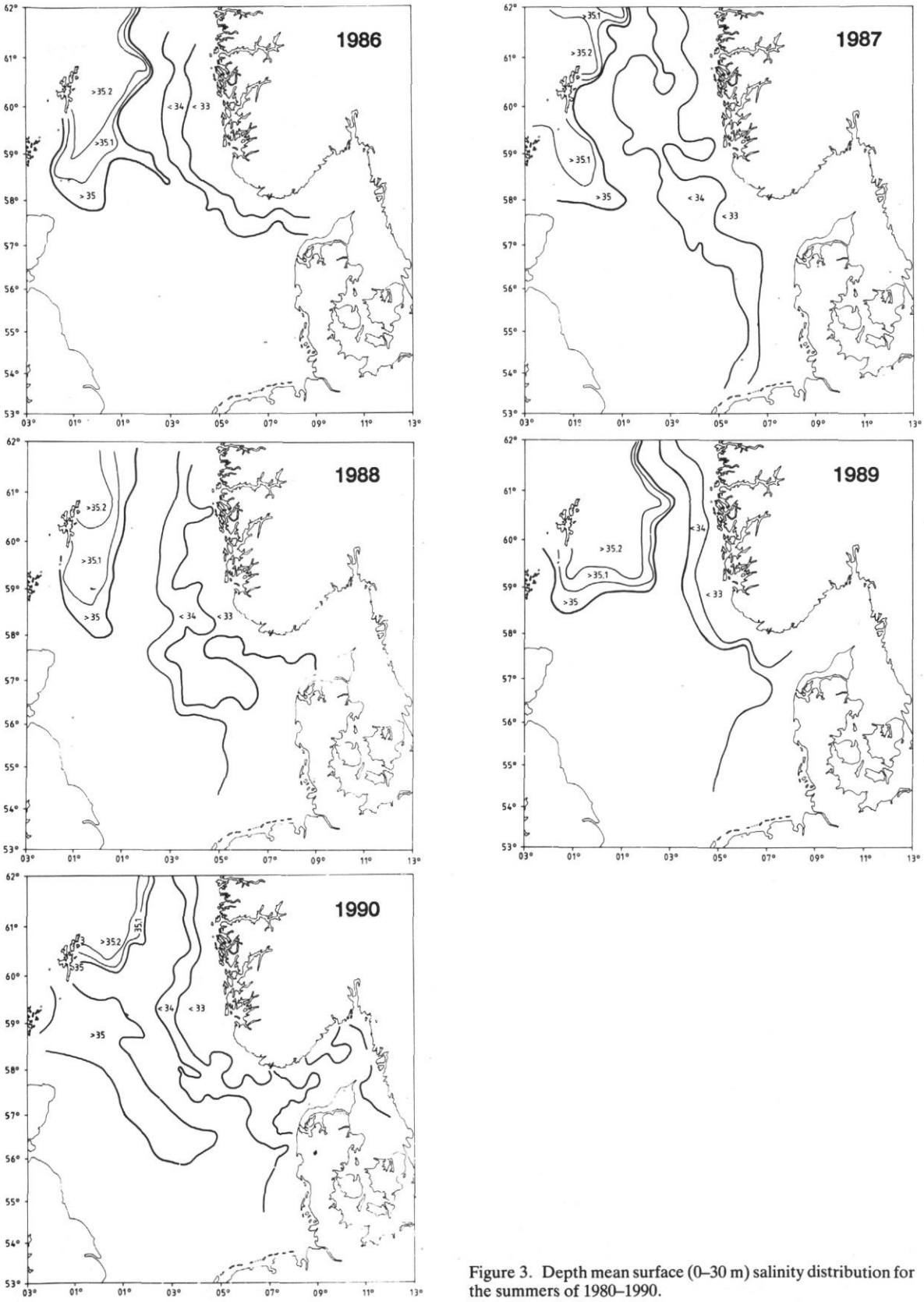


Figure 3. Depth mean surface (0–30 m) salinity distribution for the summers of 1980–1990.

mild winters we have experienced, especially in 1989 and 1990.

Figure 3 shows both the surface spread of NCW ($S < 34$ and <33) and the very strong indication of inflowing AW east of Shetland (AW_{surf}). Examples of the variable spread of the NCW are seen in the period 1986 to 1989, going from a minimum in 1986 to an extreme maximum in 1987 and back to a minimum in 1989. The spread of the near surface AW shows variability especially in its distribution pattern; however, this is to a certain extent caused by NCW sometimes reducing the surface salinity as far west as Shetland (e.g. in 1987).

Since the above parameters are partly the results of both local wind stresses in the North Sea and larger scale winds in the North Atlantic regulating the flux of AW, we at first only treat the local wind as a non-directional turbulence creator. Thus the cubed windspeed (wind effect, EFF, watt/m^2) taken every 6 h is averaged for the period April to June (here called spring or 2nd quarter), the period assumed to be most critical for the production of plankton in relation to fish. However, for the discussion of the causes of the variability of the above physical ocean parameters, the monthly mean wind stress components and wind effects from October to June are also considered. It is questionable whether these one-point wind measurements are representative for the area under consideration. However, we believe that the time averaging to a large extent also represents realistic space averages, with some restrictions due to topographic effects from Norway.

The last but not the least important ecological factor treated here is the light conditions/surface heat flux regulator, parameterized by the monthly/seasonal mean cloud cover of the low and medium clouds (CLM).

Fish-recruitment/spawning stock data are taken from the ICES VPA (Virtual Population Analysis) estimates (Anon., 1991).

Climatic time series

Hydrography

Time series of the four above-mentioned oceanographical parameters (AW_{deep}, AW_{heat}, NCW_{34}, and AW_{surf}) are shown in Figure 4a–d. Very clearly demonstrated is the well known drastic reduction in the late 1970s of the AW inflow (Fig. 4a). This figure shows three different areal coverages of AW_{deep} based on water with salinities greater than 35.0, 35.1, and 35.2. Probably the 35.1 line best represents the variability of Atlantic inflow, which reached its minimum in 1978. Since then the inflow has been relatively low, except for 1990, which is comparable to the inflow in 1971–1972.

In Figure 4b the variability of AW_{heat} is also shown for three different temperature limits, that is, for water warmer than 6.5, 7.0, and 7.5°C, being somewhat correlated with the salinity; however, there are clear discrepancies mainly due to variable winter cooling. It is interesting to note that 1990 is by far the warmest year during this whole period, and probably the warmest during the last 50 years. This is based on IMR's long hydrographical time series along the coast of Norway (unpublished data).

The spread of the NCW with salinity less than 34 (and 33) (Fig. 4c) also varies drastically from year to year. The cause has earlier been shown by Sætre *et al.* (1988) to be correlated with intense and/or persistent northerly winds causing an Ekman drift of the NCW westward in the North Sea. It is seen that such a stable surface layer might vary to cover from only about 15% to about 55% of the whole central and northern North Sea. Clearly this might be an important factor especially for the total primary production, and also for the behaviour of fish eggs and larvae. It is interesting to note that the maximum in 1987 corresponds to maximum estimates of Ceratium from the British Continuous Plankton Recording (CPR) program (Dickson *et al.*, 1991).

The last derived parameter from the hydrographic data is the areal coverage of AW at the surface (AW_{surf}) for salinities greater than 35.0 and 35.1 (Fig. 4d). After the sudden reduction from 1976 to 1978 (related again to the "Late 1970s Anomaly"), it has stayed relatively low.

The curves $S > 35.1$, $T > 6.5$, $S < 34$ and $S > 35.0$ in Figure 4a–d respectively are taken as our main hydrographic climate variables, and one can easily see that none of these are strongly correlated.

Wind and cloud cover

The wind effect (wind stress × wind speed) is anticipated to be especially important during spring for the mixing up of nutrients from the rich AW into the euphotic zone, thereby enhancing the total productivity of the ocean. Sundby and Fossum (1990) have also shown that increasing turbulence up to a certain amount increases the feeding rate of cod larvae. Figure 5a shows the average "spring" (April–June) and "winter" (October–March) non-directional wind effect. Clearly seen is the pronounced general increase, especially during winter, with an absolute maximum in 1990 and the minimum period in the late 1970s with an absolute minimum in 1978 for both "seasons". During the 1980s, the spring wind effect was relatively constant (a maximum in 1982), while large year-to-year variations are seen for the winter period.

Several authors have indicated connections between biological variations and wind directions (Colebrook, 1985; Johannesen and Tveite, 1988), and there is no doubt that the local wind direction and strength are important especially for the near surface circulation (Sætre *et al.*, 1988; Turrell *et al.*, 1991; Riepma, 1980). Therefore also the wind stress components (STRESSN,

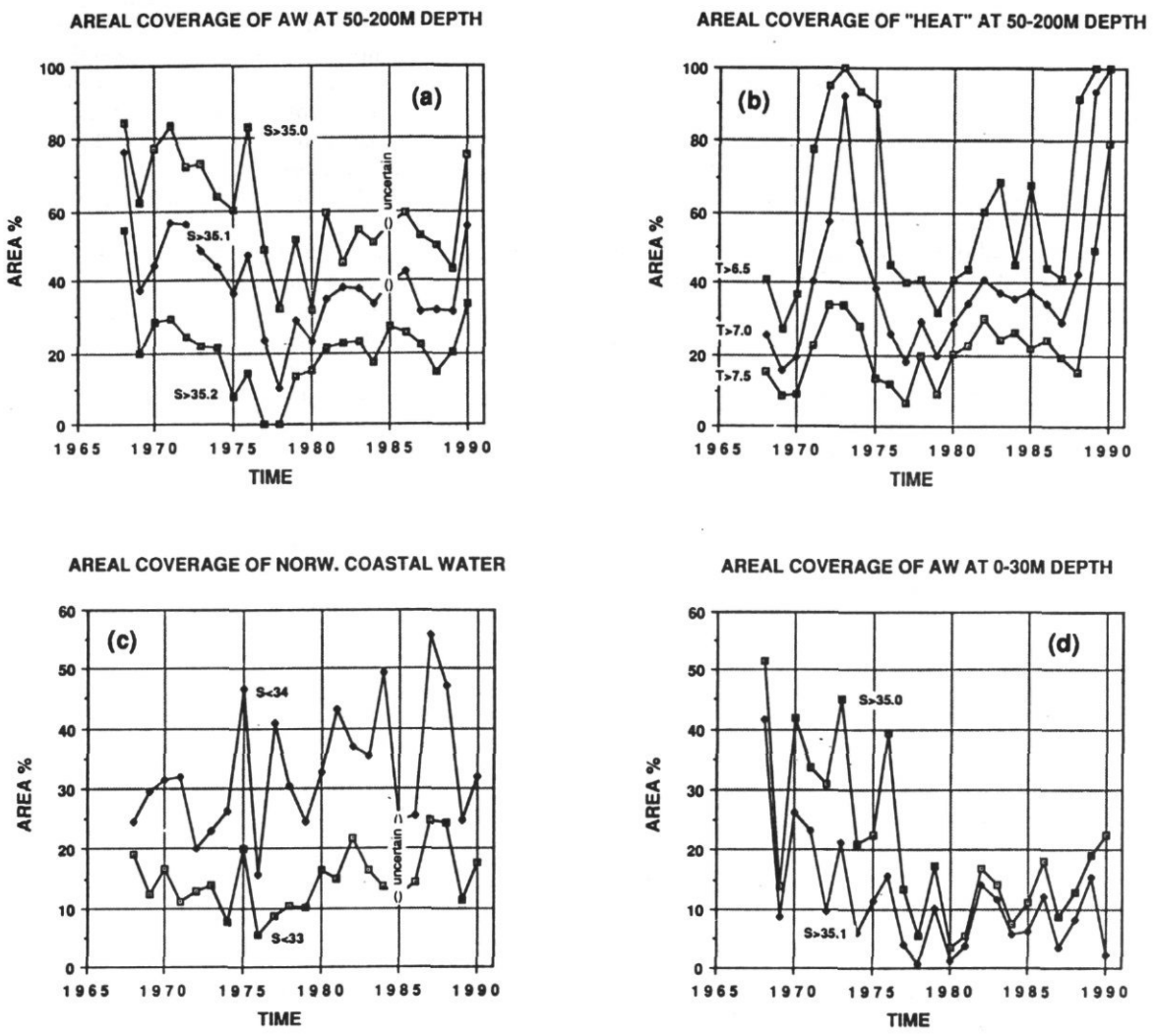


Figure 4. Time series of the fraction of the area north of 57°N covered with (a) mean salinity >35.0, 35.1, and 35.2 (AW_{deep}, ref. Fig. 1), (b) mean temperature >6.5, 7.0, and 7.5°C (AW_{heat}, ref. Fig. 2), (c) mean salinity <34.0 and 33.0 ($NCW_{34,33}$, ref. Fig. 3), and (d) mean salinity >35.1 and 35.0 (AW_{surf}, ref. Fig. 3).

positive *towards* north, and STRESSE positive *towards* east) are presented for the same seasons as above (Fig. 5b, c). During winter there is a long-term increase in both components, with large year-to-year variations. During spring the east component in general shows minor variability, while the north component indicates an enhanced variability in the 1980s compared to the 1970s. Special strong northerly (from north) winds occurred in 1981, 1982, and 1988. The combination of wind components into directions shows that the mean wind direction during winter is towards 30° (from SSW) with typical year to year variations of ±30°. The general spring wind direction is from NNW, with some extreme abnormal years, such as 1972 and 1986 with winds from SE, and 1968, 1979, (1987), 1989, 1990 with winds from between WSW and SSW.

The mean "spring" and "winter" cloud cover is presented in Fig. 5d. 1989 and 1990 had clearly the highest mean winter cloud cover for this 22-year period. The greenhouse effect that these clouds represent is probably the main cause of the subsurface NS climate in 1990 being the warmest for the last 50 years.

For some of the following statistical treatments we have used the monthly averaged wind stress components and cloud covers in addition to the "seasonal" averages presented here.

Model

When forecasting the North Sea ocean climate longer than a few weeks ahead, it is of great importance to

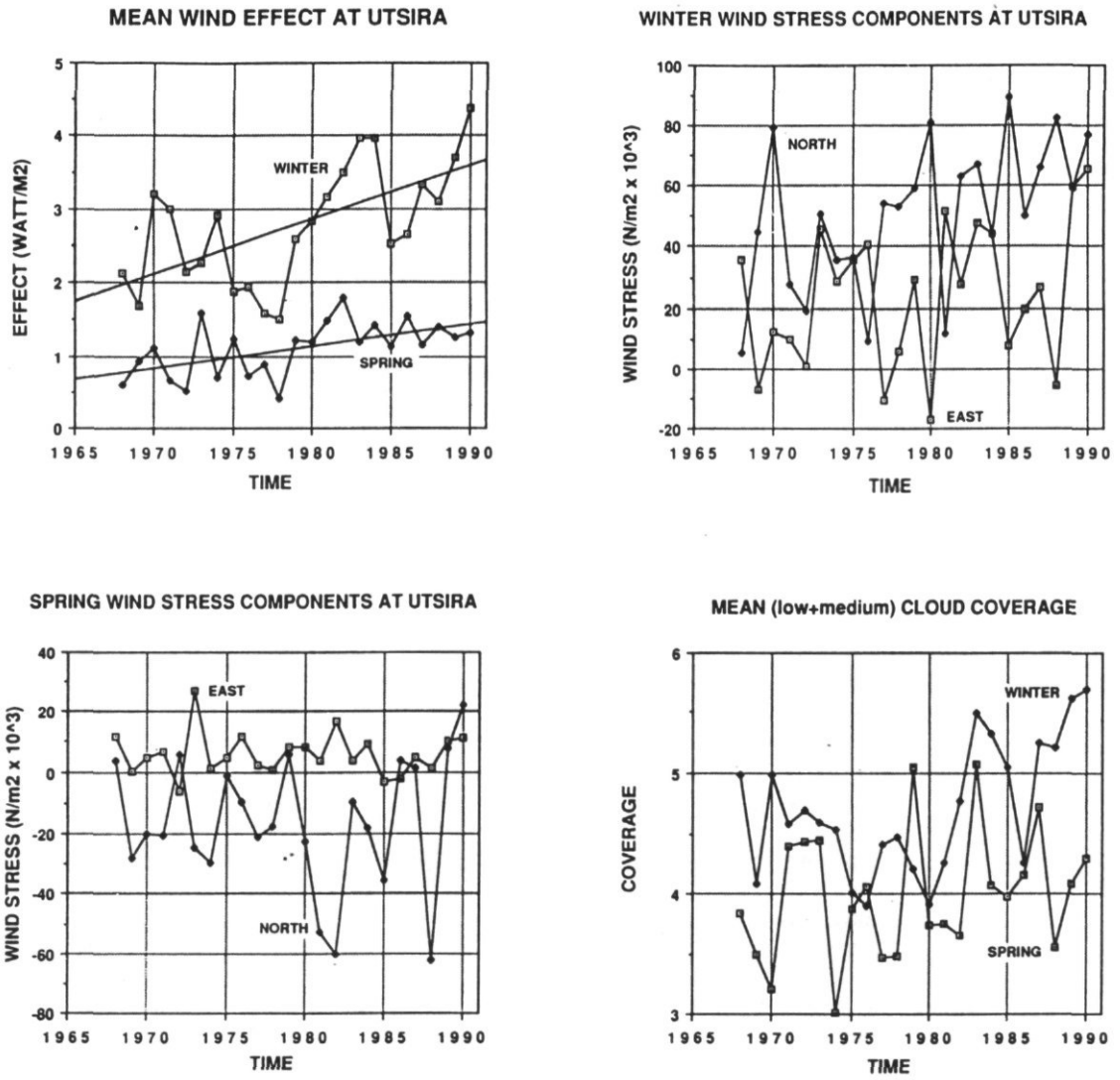


Figure 5. (a) Mean wind effect (cubed wind speed × 0.002 (=density of air × drag coefficient) for spring (April–June) and winter (October–March), (b) mean winter wind stress components (squared wind vector × 0.002) towards north and east, (c) mean spring wind stress components, and (d) mean spring and winter coverage of low and medium clouds.

evaluate the effect of local weather (especially wind) on the hydrographic variability. If strong correlations exist (and assuming that a relatively small ocean like the North Sea is not a prime steering factor of the wind climate), then there is little likelihood of predicting the variability of the ocean, since the meteorology at present can only be predicted a few days ahead.

Therefore we have run multiple linear regression analyses (Wilkinson, 1989) for each of our prime ocean climatological parameters, with our wind series and some of the other prime parameters as independent variables. In short, this regression analysis gives the coefficients a, b, c . . . from measurements in the equation:

$$\begin{aligned}\text{Ocean Variable} = \text{Const.} &+ a(AW)_{deep} + b(AW)_{heat} \\ &+ c(NCW_{34}) + d(AW)_{surf} + \\ &e_1(STRESSN_1) + e_2(STRESSN_2) + \\ &\ldots f_1(STRESSE_1) + f_2(TRESSE_2) \\ &+ \ldots + g_1(EFF_1) + g_2(EFF_2) \\ &+ \ldots + h_1(CLM_1) + h_2(CLM_2) \\ &+ \ldots \qquad (1)\end{aligned}$$

where the indexes $(_{1,2\ldots})$ indicate specific months or "seasons". It must be mentioned that the statistics do not require that these variables are independent. The "Ocean Variable" is AW_{deep}, AW_{heat}, NCW_{34}, or AW_{surf}, being respectively deleted from the right side of the equation.

The requirements we set for the selection of independent variables is that each coefficient shall be significantly different from zero, with a 95% confidence level (preferably better), and that the number of variables shall be as few as possible to reach a total *squared* multiple correlation coefficient preferably above 0.7. It is also a very important requirement that the results shall be reasonably explainable in relation to common oceanographical/biological knowledge. The use of this type of correlation analysis can of course be discussed, and many scientists will call it no more than a simple curve-fitting technique. However, knowing that the variability of certain environmental parameters is in general a combined function of *several* and not one variable, some kind of multivariate analysis is required to estimate the significance of each variable. We do not claim that the presently used statistical method is necessarily the best for our purpose; however, it seems to give interesting results.

Model results

The parameter we believe best represents the main subsurface inflow of AW to the North Sea, namely the areal spread of the mean (50–200 m) salinity greater than 35.1 (AW_{deep}, Fig. 4a), is not well enough explained by our other variables. This means that the pulsation of the AW outside the North Sea (probably through the Faroe–Shetland channel, FSC) seems to have a significant influence on the transport into the North Sea. This pulsation is more connected to the large-scale winds over the North Atlantic, which are not represented by the one-point wind measurements at Utsira.

The variation of "heat" in the North Sea represented by the areal coverage of mean (50–200 m) temperature above 6.5°C (Fig. 4b), was successfully modelled as shown in Figure 6. The obtained model equation is:

$$\begin{aligned} MODAW_{heat} = {} & -63.31 + 0.85 \times (STRESSE_{Des}) + \\ & 0.56 \times (STRESSN_{Jan}) + 4.649 \times \\ & (CL_{Mar}) + 16.44 \times (CL_{Apr}) - 0.43 \times \\ & (STRESSN_{Apr}) \end{aligned} \quad (2)$$

One must be careful interpreting this result, especially since the number of independent variables is as large as 5. (Clearly, the air–sea temperature difference and the air humidity are important parameters (not considered) for the sea surface heat flux and thereby the AW_{heat} parameter.) However, to a certain degree the five variables can be reduced to three explanatory processes such that: A warm (versus cold) ocean seems to be the result of a combination of: (1) strong/persistent eastward wind stress in December turning to strong/persistent northward wind stress in January. This increases the general circulation and surface outflow, giving room for subsurface inflow of relatively warm AW, and brings relatively warm air from the south and/or west; (2) heavy cloud cover in March and April reducing the radiative cooling; and (3) strong southward (or reduced northward) wind stress in April. The importance of high cloud cover, especially in April, and the southward wind stress, causing westward spreading of light NCW, will reduce the chances of a total breakdown of the vertical stability, thus reducing the possibility of the cold surface water being mixed down into the deep water.

We had expected that the AW_{deep} would be an explanatory factor for the heat variability, but at this stage it seems that the local atmospheric effects are more important than the variability in the inflowing AW. This cause–effect study of the heat parameter seems to include two distinct time scales: one due to winds about six months prior to observations, and one due to winds and clouds about three months before.

As qualitatively stated by Sætre *et al.* (1988), we find that the large (versus small) summer areal spread of the NCW_{34} (stable surface layer) measured around June/July (Fig. 4c) is well quantified (empirical model, Fig. 7), and mainly caused by (1) strong/persistent southward wind stress components in May and June causing a westward Ekman drift of this surface water. The obtained model equation is:

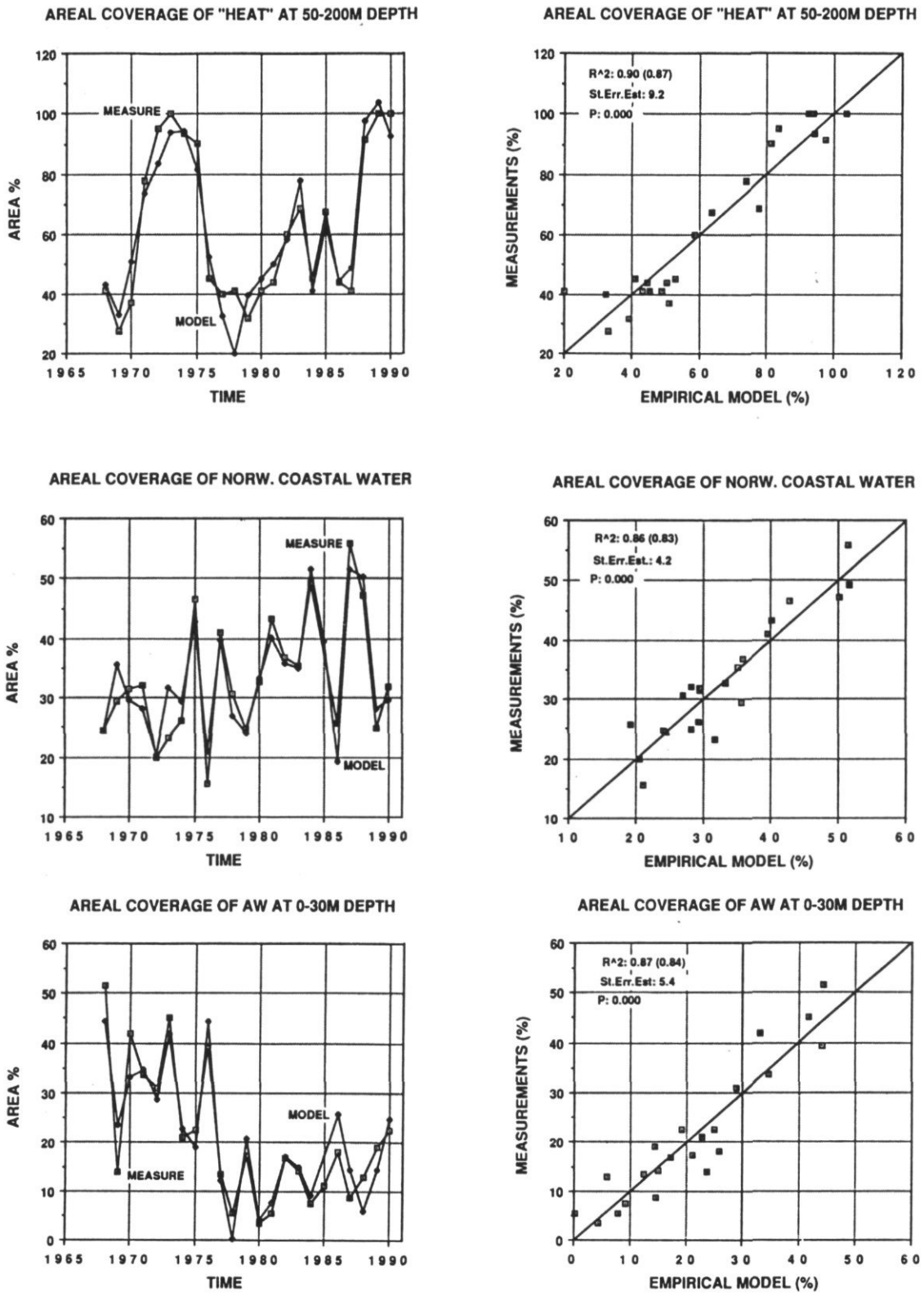
$$\begin{aligned} MODNCW_{34} = {} & 21.60 - 0.25 \times (STRESSN_{May}) - 0.29 \\ & \times (STRESSN_{June}) + 0.18(EFF_{Spring}) \\ & - 0.09 \times (STRESSSN_{Feb}) \end{aligned} \quad (3)$$

Additional causes for large spread are given by (2) strong mean spring wind effect (Fig. 5a) (large total energy input resulting in increased meandering and eddy activity) and strong southward (or weak northward) wind stress in February. The effect of the wind stress in February might be explained by its causing increased westward drift of relatively fresh water in the southern North Sea, influencing the northern North Sea about four months later. Because of very limited data coverage in 1985, this year is excluded in this analysis.

The variability of the areal coverage of AW at the surface (AW_{surf}) is well quantified (empirical model, Fig. 8) by a combination mainly of (1) areal coverage of AW with salinity greater than 35.0 (Fig. 4a) and (2) strong/persistent eastward wind component in May. This means that the surface inflow of AW is, as expected, correlated with the "deep" AW inflow and westerly winds causing a direct Ekman drift southwards. The model equation is:

$$\begin{aligned} MODAW_{surf} = {} & -9.21 + 0.72(AW_{deep}) + 0.84 \times \\ & (STRESSE_{May}) - 0.25 \times (NCW_{34}) - \\ & 0.19 \times (AW_{heat}) \end{aligned} \quad (4)$$

An additional contribution to AW_{surf} is (3) reduction of the areal coverage of NCW (overriding the AW_{surf}), but more strangely, (4) reduction of the areal coverage of



Figures 6, 7, and 8. Comparisons between measurements and empirical models for AW_{heat} (Fig. 6, top), NCW_{34} (Fig. 7, centre), and AW_{surf} (Fig. 8, bottom).

the "deep" water (AW_{heat}) with mean temperature greater than 7.5°C (Fig. 4b). According to Steele (1957) the persistence of cold, dense bottom water over the northern North Sea plateau will prevent Atlantic inflow to the area during summer. This seems not to be correct (as also stated by Svendsen *et al.*, 1991), but these results show that this effect might hinder the deeper inflow of AW, being lifted to flow southward near the surface.

To prove that these suggested empirical models are realistic and not just "curve fitting", we should have hidden about 20% of the dependent variables (say the last four years), used the already selected independent variables and run the regression analysis again to re-estimate the coefficients. The difference between the new model results and the hidden data would then roughly be within the Standard Error of Estimates given in each of the scatter plots in Figures 6, 7, and 8. At present we have left this exercise in relation to the empirical models of the biological variables (Svendsen *et al.*, 1991).

An example demonstrating the climatic importance for the recruitment success or failure of 0-group whiting (Anon., 1991) is demonstrated in Figure 9, using the same regression technique as above. The empirical model equation is:

$$MODWHIT(0) = 57.72 + 0.52(AW_{heat}) - 36.69(EFF_{2.QU}) \quad (5)$$

This shows that good recruitment is obtained after a mild winter and/or strong inflow of AW, and relatively weak wind conditions in April, May, June (2nd quarter). More than 70% of the recruitment variability in the period 1970–1988 is explained by these two climatological parameters alone.

Conclusion

From routine hydrographic surveys in the North Sea during the last 22 years, we have extracted parameters which we believe better describe the climate variability than the observations (temperature and salinity) themselves.

The spread of high salinity water (AW_{deep}) is a good indicator of the previous (maybe half year) inflow of AW, showing a minimum in 1978 related to "the Late 1970s anomaly", and a maximum in 1990 comparable to the inflow in 1971–1972. Measured local winds at western Norway (Utsira) cannot significantly explain these variabilities. That the main inflow of AW to the North Sea is not clearly correlated with the local wind gives hope that its effect on the North Sea can be predicted at least a half to one year ahead if measured or realistically modelled, e.g. in the Faroe–Shetland Channel (FSC).

The spread of "heat" (AW_{heat}) is a combined indicator of the previous inflow of AW and winter cooling.

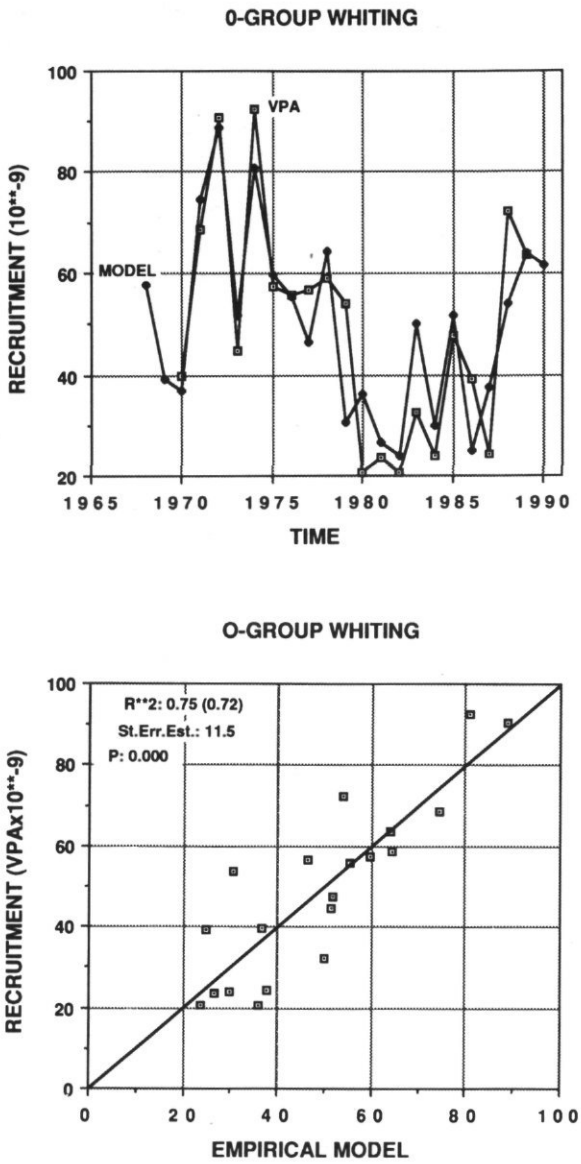


Figure 9. Comparisons (annual time series and scatter plot) between measurements (VPA) and empirical model for the 0-group recruitment of whiting.

The variability is quantified by a combination of monthly mean wind stress and cloudiness parameters (Eq. (2), Fig. 6); 1990 is by far the warmest year during this whole period, probably the warmest during the last 50 years.

The spread of a stable surface layer (NCW_{34}) is clearly correlated with northerly winds during May and June with a clear maximum in 1987 (Eq. (3), Fig. 7).

A significant near surface AW inflow spreading in the northwestern North Sea (AW_{surf}) is clearly demonstrated (Fig. 3). The variability is mainly related to the main AW inflow (AW_{deep}) and westerly winds (Eq. (4), Fig. 8).

The time series of these physical oceanic parameters together with the wind climate are used primarily to explain significant parts of the variability of the recruitment of whiting. The climatic effects on some other gadoid stocks, herring, and the fraction of western mackerel entering the North Sea is further discussed by Svendsen *et al.* (1991). Some of these parameters seem, in combination with meteorological variables, to be of prime importance for the recruitment success (or not) partly indirectly through their effect on primary and secondary production. It is open to discussion if the extracted parameters are the best choice. However, this is a first (and successful) try at using an extensive amount of collected data to *quantify* the integrated effect of several (hopefully the most) important climatic factors on the marine life.

Even if we cannot explain the full coupling from the physics, through chemistry, primary, and secondary production, fish spawning, areal distribution and predation, to the recruitment success, the results indicate that the climate has an effect on most (maybe all) stages. It also indicates that using just one single climatic parameter to describe biological variability will in general lead to failure.

For the future it will be interesting to see if some of our oceanographic variables can be explained by one or a few of our long-term monitoring stations, reducing the need for yearly large ship expeditions, and giving the opportunity to study a period of more than 50 years. Some sudden climatic variations in the FSC seem to give maximum integrated response approximately two years later in the North Sea (unpublished results) and about three years later in the Barents Sea (Dickson and Blindheim, 1984). This indicates the importance of regular monitoring of the FSC for possible forecasting of some of the climatic variabilities in these oceans, especially of the inflowing AW.

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Nutrients in the German Bight: concentrations and trends

Dieter Körner and Günter Weichart

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In January/February of the years 1978, 1985, 1987, 1989 and 1990 nutrient concentrations were measured in the German Bight and compared with data set from January/February 1936. In this mixing area of fresh water with North Sea water, the nutrient distribution during winter time shows a good linear nutrient–salinity relation. By using regression lines the nutrient concentrations of different years can be compared in spite of variations due to the actual hydrographic situation and fresh water runoff. Compared with 1936, the phosphate concentrations in near-coastal areas have greatly increased, while silicate concentrations have increased only little. Nitrate concentrations were not measured in 1936. Since 1978, the measurements do not show any further increase of the nutrient concentrations in the German Bight.

Dieter Körner and Günter Weichart: Federal Maritime and Hydrographic Agency, Bernhard-Nocht-Strasse 78, D–2000 Hamburg 36, Germany.

Introduction

The distribution of nutrient concentrations in the German Bight is determined not just by the mixture of nutrient-rich fresh water and North Sea water, but also by the biochemical activity of plankton, bacteria, etc., and by the actual hydrographic conditions. To identify trends, the frequency of measurements must either be greatly increased or measurement conditions and evaluation methods must be such that the variability caused by biochemical and hydrographic parameters does not falsify the results.

Biochemical activity is at its lowest in the winter months. Even in exceptionally mild winters, e.g. 1989 and 1990, the near-surface and near-bottom pH values of 8.0 to 8.1 for the entire German Bight show biochemical metabolism to be low. This is why we only evaluate winter nutrient concentrations for a trend analysis. Salinity normalization makes allowances for hydrographically induced differences in nutrient distribution.

As early as January/February 1936, Kalle (1937) measured nutrient concentrations in the entire German Bight. Encouraged by the beginning of discussion about eutrophication, Weichart (1986) carried out nutrient measurements in January 1978. More measurements followed at the end of January/beginning of February 1985; at the end of January 1987; at the beginning of February 1989, and in January 1990. The results are evaluated comparatively in this study.

Methods

Water samples were taken at various depths using plastic Nansen bottles (Hydrobios, Kiel) or a rosette sampler with Niskin bottles (General Oceanics, Miami). The salinity of the samples was measured using an inductive salinometer.

The following nutrient analyses were carried out on board using a filterphotometer (Elko II, Carl Zeiss, Oberkochen): Dissolved inorganic phosphate according to Murphy and Riley (1962); precision (95% confidence range): $\pm 0.03\,\mu$mol/l. Ammonia nitrogen according to Koroleff (1983); precision (95% confidence range): ±8%. Nitrite nitrogen according to Grasshoff (1983); precision (95% confidence range): $\pm 0.2\,\mu$mol/l.

Samples of silicate, nitrate+nitrite, total nitrogen, total phosphate and, from January 1990, also of dissolved inorganic phosphate were deep-frozen and analysed immediately on return: Silicate and nitrate+nitrite in the autoanalyser in line with Technikon's instructions (1972); total nitrogen after oxidation in an alkaline medium; precision (95% confidence range): $\pm 0.3\,\mu$mol/l for silicate, $\pm 0.5\,\mu$mol/l for nitrate+nitrite, ±7% for total nitrogen. Up to 1989, total phosphate was determined photometrically after digestion with potassium peroxodisulphate according to Koroleff (1968); precision (95% confidence range): $\pm 0.05\,\mu$mol/l. After January 1990, total phosphate, like dissolved phosphate, was determined in the autoanalyser according to

Mangelsdorf (1972); precision (95% confidence range): ±3%.

The phosphate, silicate, nitrite and nitrate+nitrite samples were filtered before analysis or deep-freezing. The total phosphate and total nitrogen samples were filtered after digestion. To prevent contamination, the ammonium samples were not filtered. Instead, their turbidity blanks were measured as part of the photometric analysis and subtracted from the results. The same procedure was applied to the 1978 samples for phosphate.

The samples analysed in 1936 were filtered as described above. Weichart (1990) examined the comparability of the old phosphate data with the new.

Nutrient concentrations in January/February 1978, 1985, 1987, 1989, and 1990

The high nutrient input via rivers is responsible for the concentration pattern found in the German Bight: higher concentrations with low salinity near the coast; lower concentrations in water with a greater salinity (Figs. 1–2). The concentration patterns for total phosphate, ammonium, and total nitrogen (not shown here) are similar. The complete sets of distribution charts of the years mentioned above including the results from January/February 1936 are published in *Meereskundliche Beobachtungen und Ergebnisse* (1992).

Nitrite distribution, however, does not match the above distribution pattern (Fig. 3): besides elevated values near the coast, a secondary maximum occurs near the convergence zone. Kalle drew attention to this as early as 1937.

Large or small differences are seen in the salinity and nutrient distribution in individual areas, depending on medium-term weather conditions. In mild winter months when winds mainly come from the southwest, such as in January/February 1989 and 1990, a very distinct freshwater plume with a high concentration of nutrients moves along the coast of Schleswig-Holstein. In cold winter months when winds mainly come from the east, such as in January/February 1985 and 1987, the freshwater part here is smaller and the water contains a considerably lower concentration of nutrients.

Just how different water can be at the same site at different times is shown using the position 54°50′ off Sylt (cf. Figs. 1 and 2): in February 1989, salinity here was only 27.51. If pure North Sea water is assumed to have a salinity level of 35, this would mean a freshwater part of 21.4%. In 1987, salinity was 30.56 and in 1985 30.52, a freshwater part of only 12.7%. It is clear that such differences in the mixture ratios must result in different nutrient concentrations.

Nutrient–salinity relation

Nutrient concentrations measured at different times at the same site cannot be compared for trend analysis; only those measured in water bodies with the same proportions of salt and fresh water can be compared. For the normalization against salinity of nutrient concentrations measured in the mixed water, the linear regression procedure presents itself. Kalle (1956) used this method when estimating biologically induced nutrient metabolism in the German Bight.

The mixture lines calculated for the winter data for 1978, 1985, 1987, 1989, and 1990 are given in Figures 4–5; the accompanying regression parameters (intercept and slope with standard deviation, and correlation coefficient) are given in Tables 1–5.

Analysis shows there to be a very good linear correlation between nutrient concentrations and salinity in the salinity range examined. This is all the more remarkable in that a strong linear correlation can only be expected under certain conditions: (1) the nutrients must behave conservatively; (2) the nutrient concentrations in the various freshwater tributaries must be approximately equal; (3) major temporal changes in concentration must not occur in the mixture components – at least not in the mixing period preceding the measurement.

Ad 1

Nutrients can only be expected to be fairly conservative in the winter when biological activity is negligibly low not just in respect of nutrient consumption in the euphotic layer. The degradation of organic material in the water column and at the interface between water and sediment should also largely be completed or have ceased. The extent to which such processes affect nutrient concentrations in the winter is currently being investigated in the OPTINOM programme.

As already mentioned, nitrite does not behave conservatively. Because of its secondary concentration maximum in the convergence zone the nitrite–salinity relation is not linear (correlation coefficient of 0.21). Owing to the comparatively small nitrite concentration, this does not, however, noticeably influence evaluation of the sum parameters nitrate + nitrite or nitrate + nitrite + ammonium (compare Table 2).

Atmospheric input of nitrogen compounds cannot be ignored (see e.g. Gerlach, 1990), as it can slightly increase the concentration in the whole water column after vertical mixing. The linearity of the nutrient–salinity relation is not affected by this.

It is not to be expected that nutrients will behave conservatively in the turbidity zone, which lies well beyond our area of investigation. The values for the intercept in Tables 1–5 can nonetheless be compared with the concentrations measured in the Elbe. This surprising result only means that all nutrients trans-

Table 1. Regression parameters of the nutrient–salinity relation in the German Bight in January 1978.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	308	−0.36	0.01	13.18	0.25	−0.94
Total phosphorus	291	−0.41	0.01	15.05	0.35	−0.91
Silicate	318	−5.26	0.12	186.69	4.08	−0.92
Nitrate+nitrite	317	−10.66	0.20	376.67	6.66	−0.95

Table 2. Regression parameters of the nutrient–salinity relation in the German Bight in January/February 1985.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	188	−0.34	0.01	12.32	0.17	−0.98
Total phosphorus	178	−0.51	0.02	18.62	0.54	−0.92
Silicate	190	−4.42	0.07	157.09	2.47	−0.97
Nitrate+nitrite	189	−9.52	0.11	337.67	3.70	−0.99
Ammonium	190	−1.89	0.06	65.96	1.92	−0.92
Nitrate+nitrite+ammonium	189	−11.40	0.12	403.52	4.10	−0.99
Total nitrogen	187	−13.51	0.26	486.83	8.58	−0.97

Table 3. Regression parameters of the nutrient–salinity relation in the German Bight in January 1987.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	83	−0.26	0.01	9.77	0.33	−0.94
Total phosphorus	77	−0.47	0.03	17.03	1.09	−0.85
Silicate	83	−3.99	0.16	141.38	5.31	−0.94
Nitrate+nitrite	83	−6.81	0.21	246.17	6.81	−0.96
Ammonium	53	−1.05	0.08	36.75	2.60	−0.88
Nitrate+nitrite+ammonium	53	−7.69	0.30	277.73	9.60	−0.96
Total nitrogen	76	−11.40	0.46	408.51	14.74	−0.95

Table 4. Regression parameters of the nutrient–salinity relation in the German Bight in February 1989.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	114	−0.28	0.01	10.21	0.33	−0.93
Total phosphorus	113	−0.41	0.02	14.92	0.53	−0.92
Silicate	113	−4.73	0.18	164.48	5.79	−0.93
Nitrate+nitrite	112	−10.95	0.19	384.87	6.25	−0.98
Ammonium	101	−0.97	0.05	33.82	1.52	−0.90
Nitrate+nitrite+ammonium	99	−11.71	0.22	411.82	7.09	−0.98
Total nitrogen	112	−12.83	0.22	458.11	7.28	−0.98

Table 5. Regression parameters of the nutrient–salinity relation in the German Bight in January 1990.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	79	−0.34	0.01	12.22	0.32	−0.97
Total phosphorus	66	−0.45	0.03	16.42	0.92	−0.90
Silicate	79	−4.69	0.27	161.34	8.97	−0.89
Nitrate+nitrite	79	−7.17	0.26	252.67	8.72	−0.95
Ammonium	79	−1.44	0.12	51.71	4.13	−0.80
Nitrate+nitrite+ammonium	79	−8.61	0.22	304.37	7.47	−0.97
Total nitrogen	79	−13.61	0.43	481.79	14.22	−0.96

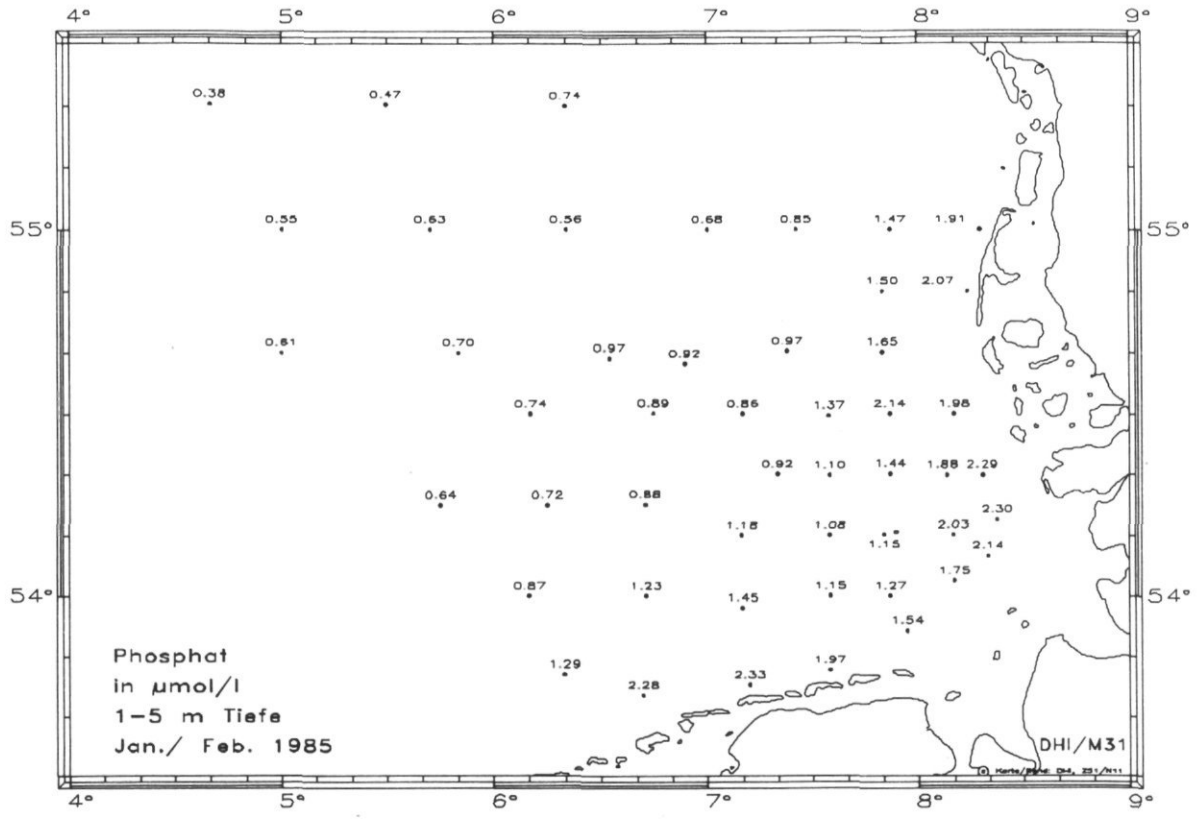

Figure 1a. Phosphate distribution in the German Bight at the end of January/beginning of February 1985.

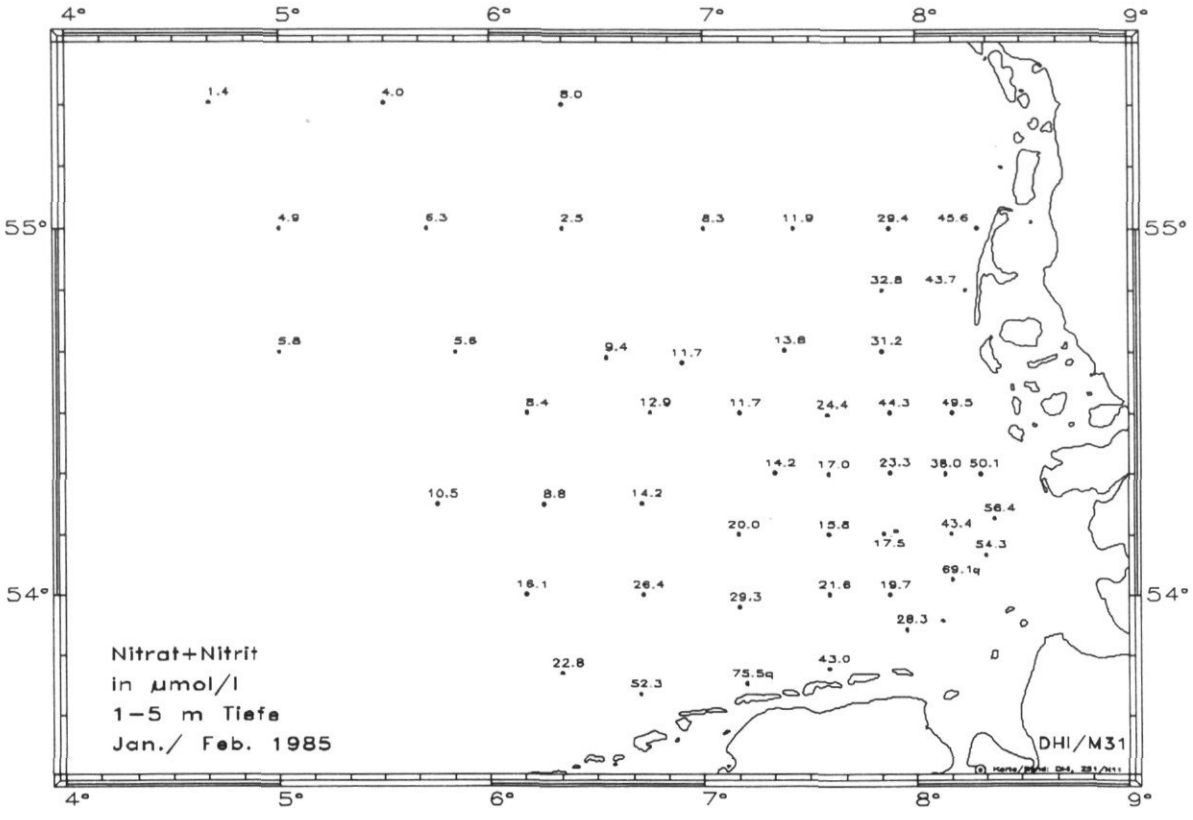

Figure 1b. Nitrate distribution in the German Bight at the end of January/beginning of February 1985.

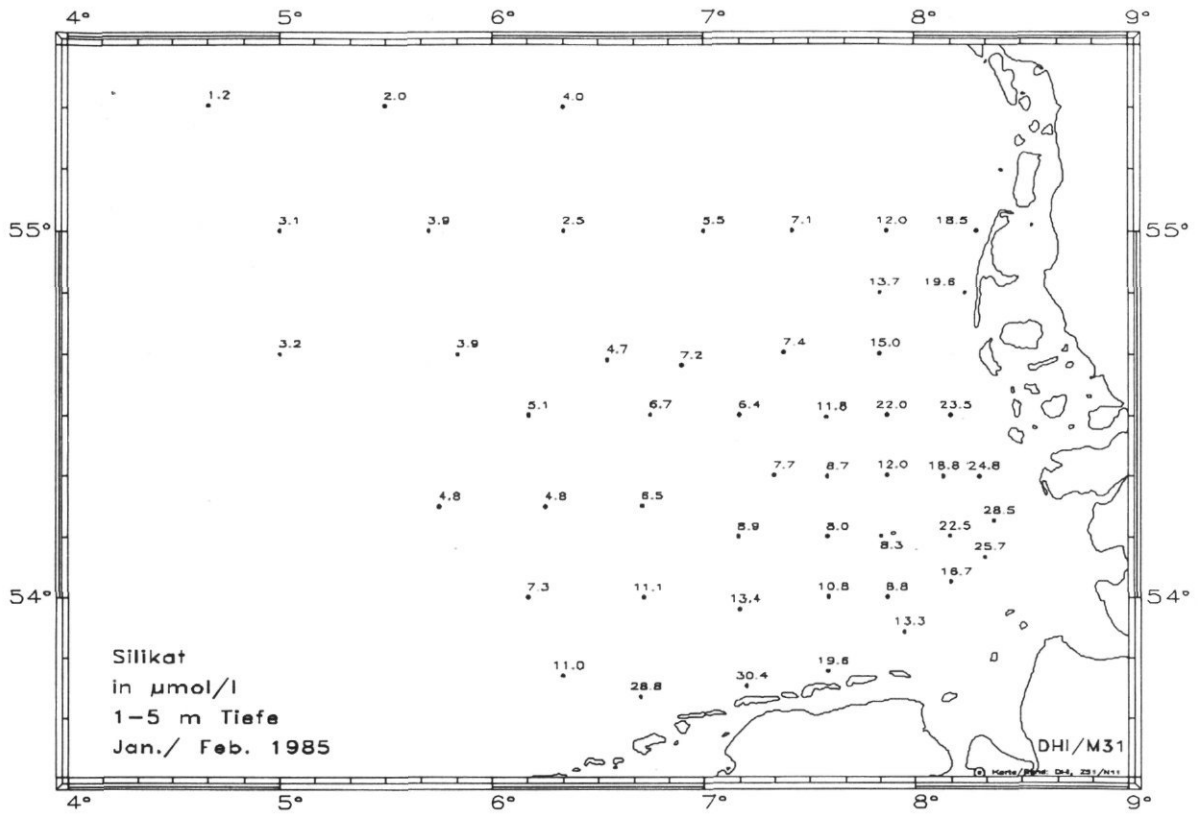

Figure 1c. Silicate distribution in the German Bight at the end of January/beginning of February 1985.

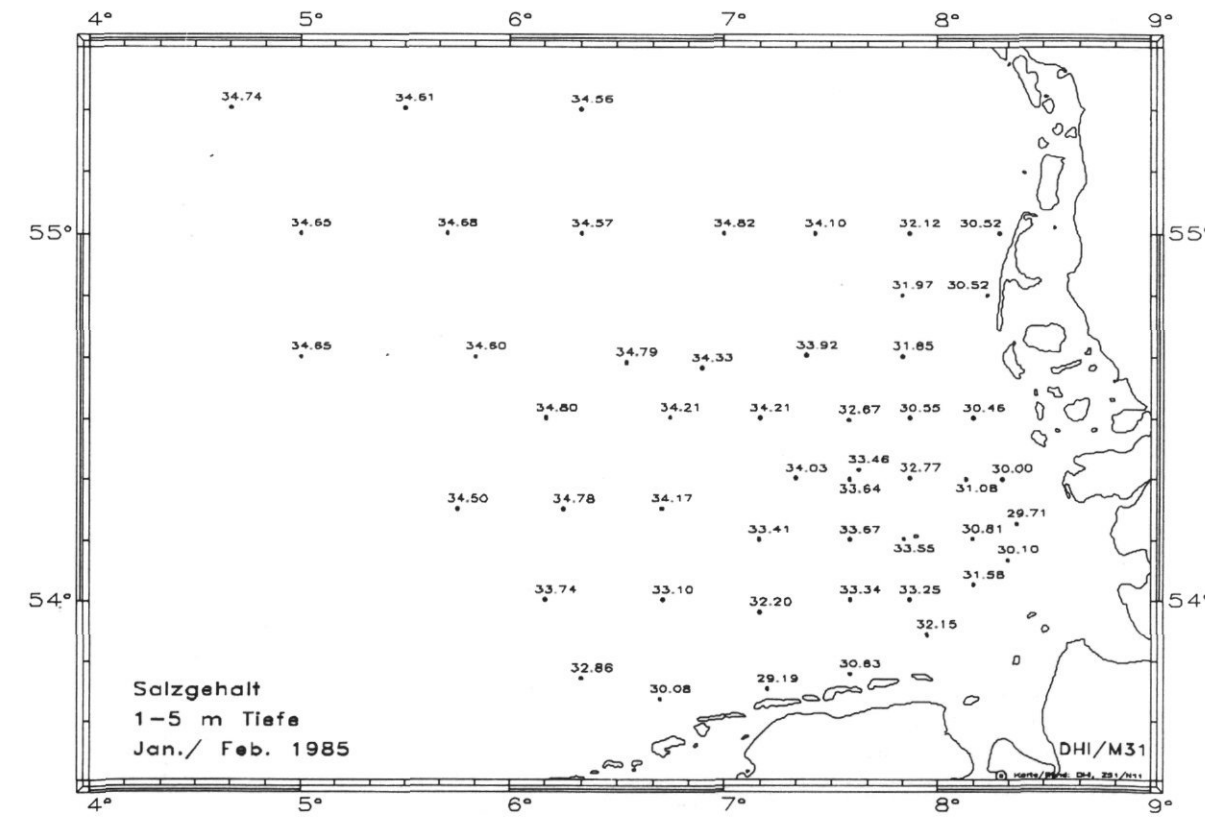

Figure 1d. Salinity distribution in the German Bight at the end of January/beginning of February 1985.

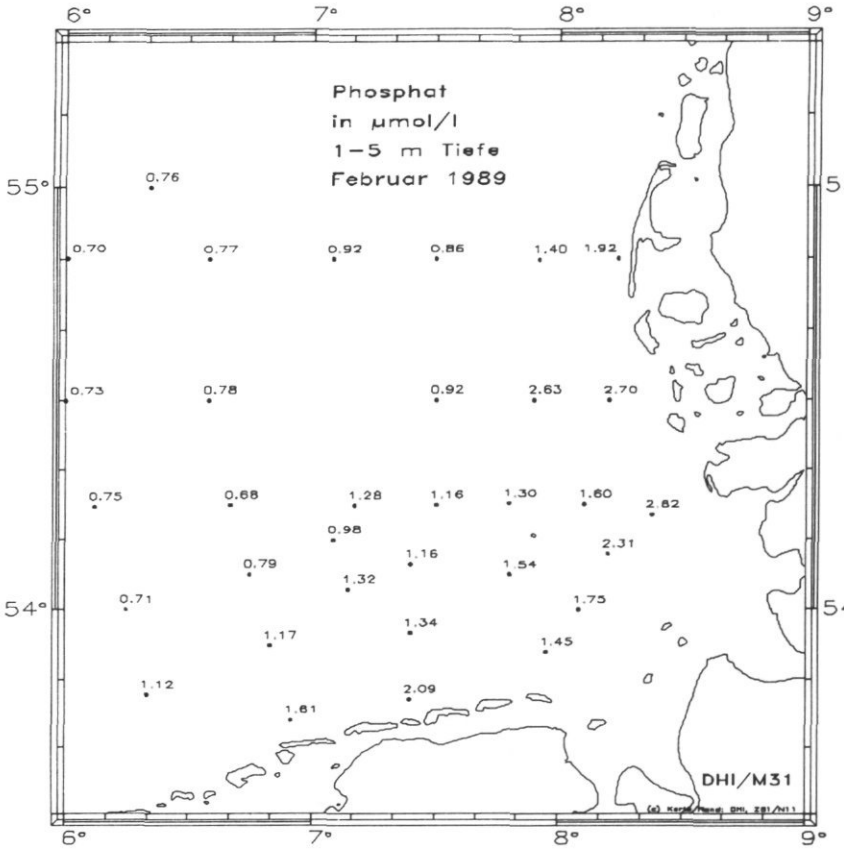

Figure 2a. Phosphate distribution in the German Bight at the beginning of February 1989.

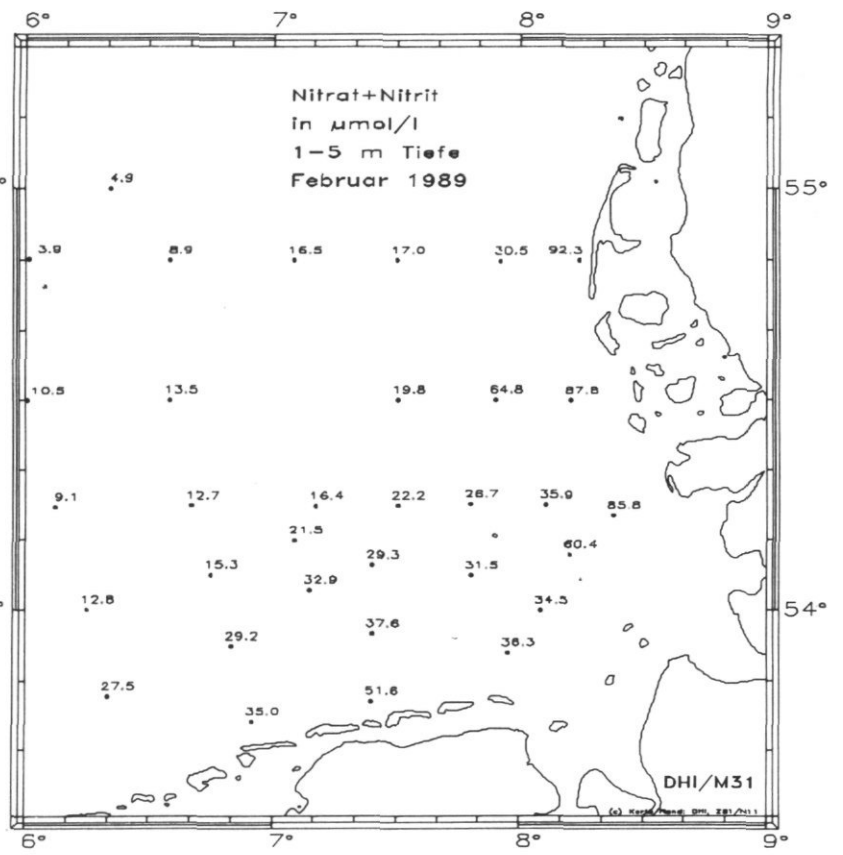

Figure 2b. Nitrate distribution in the German Bight at the beginning of February 1989.

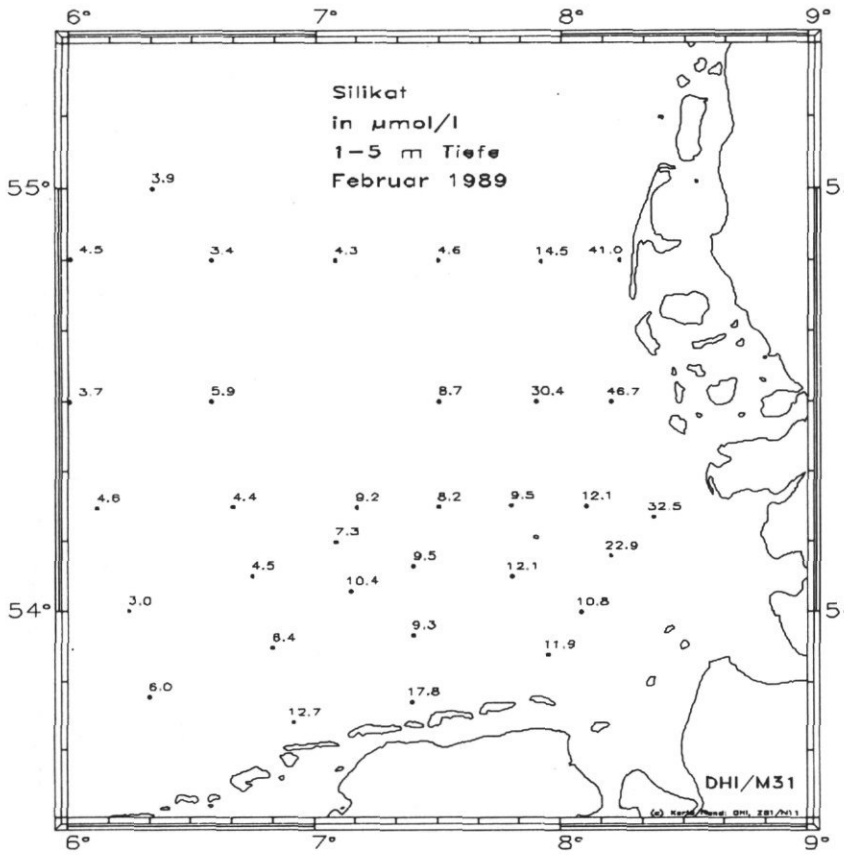

Figure 2c. Silicate distribution in the German Bight at the beginning of February 1989.

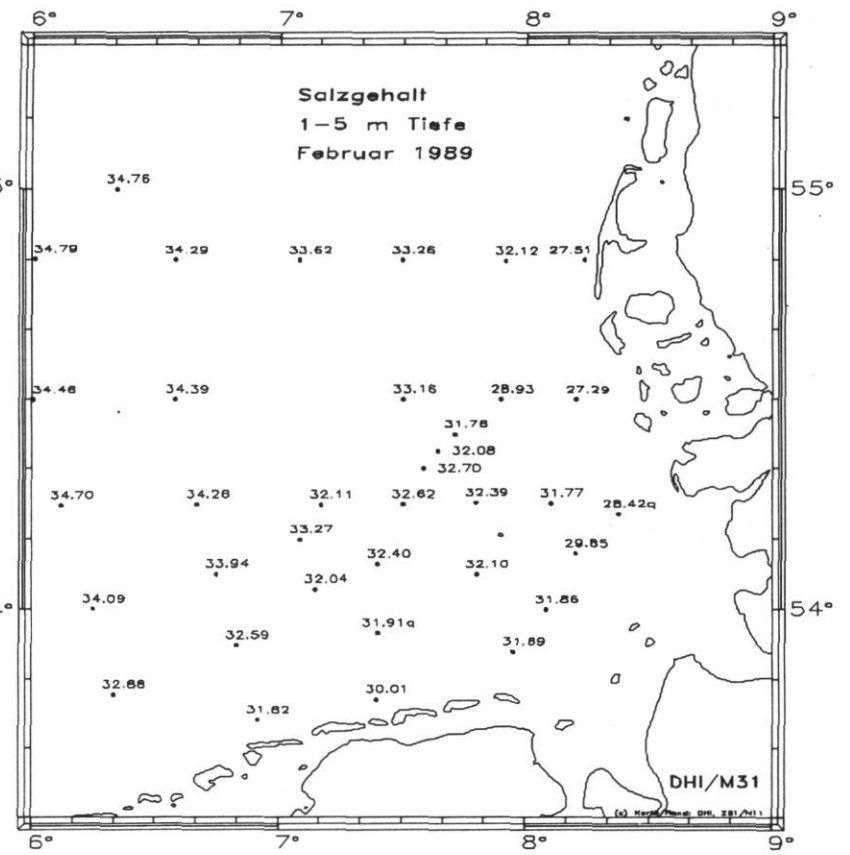

Figure 2d. Salinity distribution in the German Bight at the beginning of February 1989.

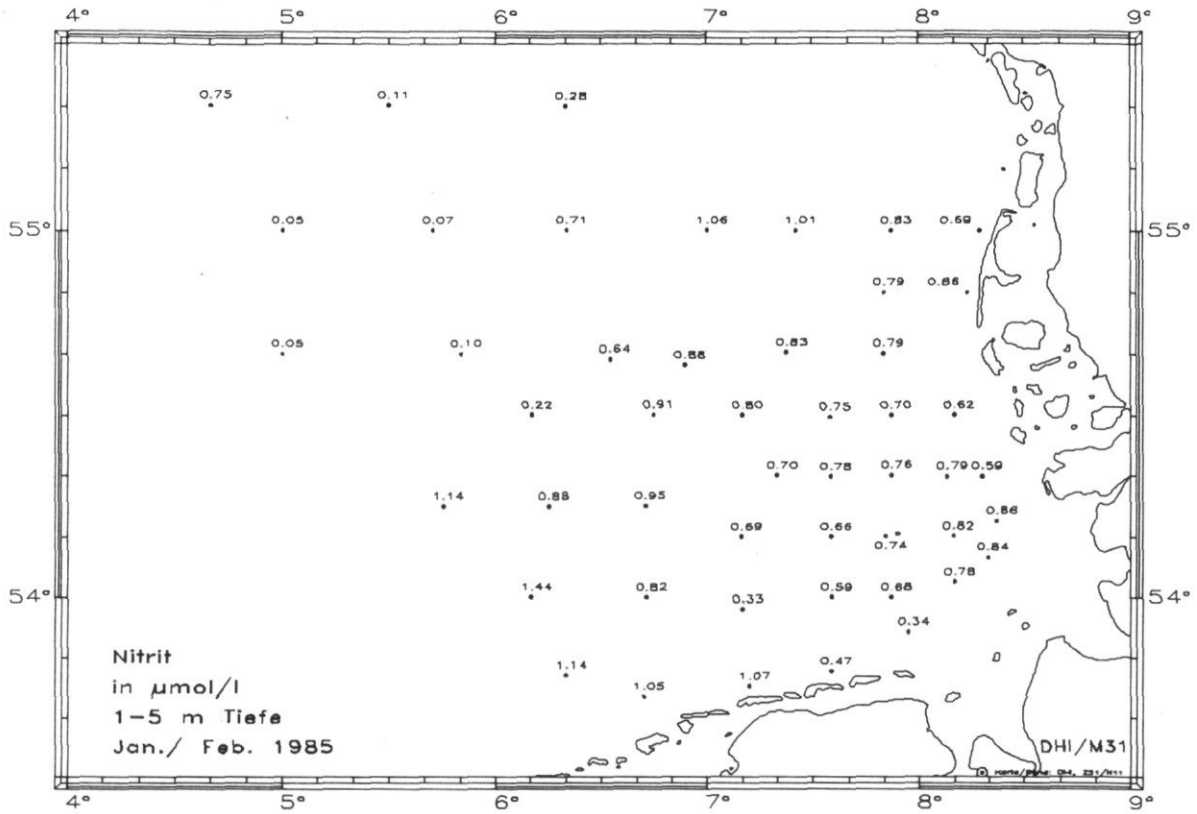

Figure 3. Nitrite concentration in the German Bight at the end of January/beginning of February 1985.

ported by freshwater tributaries reach the German Bight in spite of the complex physical and chemical processes which take place in the brackish water zone.

Ad 2 and 3

The nutrient concentrations measured in the freshwater boundary in recent years in the Elbe, Weser and Rhine (cf. tables and results by ARGE Elbe, ARGE Weser, and the International Commission for the Protection of the Rhine) are not generally very different from each other; the temporal variations are slightly greater.

Evaluation of near-surface concentrations

To obtain a higher statistical reliability the concentration values from all the sampling depths were included in the regression analysis. This is why individual "outliers", whose cause was partly found to be local pollution near the surface, do not affect the final result.

In some regression diagrams, however, concentrations frequently occur which clearly lie above the mean value as given by the regression line. Among these are total phosphate in 1985 and 1978 and phosphate in 1978. These are near-bottom results from stations in shallower water where the motion of the sea caused the sediment to swirl about, thus possibly releasing nutrients. This should above all affect total phosphate and total nitrogen concentrations. For this reason and to enable us to compare the results from 1936, when only surface samples were analysed, we evaluated the near-surface results for the winters of 1978 and 1985 separately (Tables 6–7). This improves the look of some diagrams, but the results for most nutrients are only slightly different from those in Tables 1 and 2. A signifi-

Table 6. Regression parameters of the nutrient–salinity relation in the surface waters of the German Bight in January 1978.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	52	−0.35	0.02	12.71	0.60	−0.94
Total phosphorus	47	−0.41	0.03	14.98	0.85	−0.92
Silicate	56	−5.59	0.30	197.81	9.86	−0.93
Nitrate+nitrite	56	−10.78	0.50	380.74	16.43	−0.95

Table 7. Regression parameters of the nutrient–salinity relation in the surface waters of the German Bight in January/February 1985.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	49	−0.32	0.01	11.84	0.30	−0.98
Total phosphorus	47	−0.43	0.03	16.00	0.84	−0.93
Silicate	49	−4.62	0.15	163.67	4.86	−0.98
Nitrate+nitrite	49	−9.64	0.21	341.43	6.81	−0.99
Ammonium	49	−1.90	0.11	66.32	3.50	−0.93
Nitrate+nitrite+ammonium	49	−11.54	0.24	407.76	7.88	−0.99
Total nitrogen	48	−13.01	0.46	468.44	15.20	−0.97

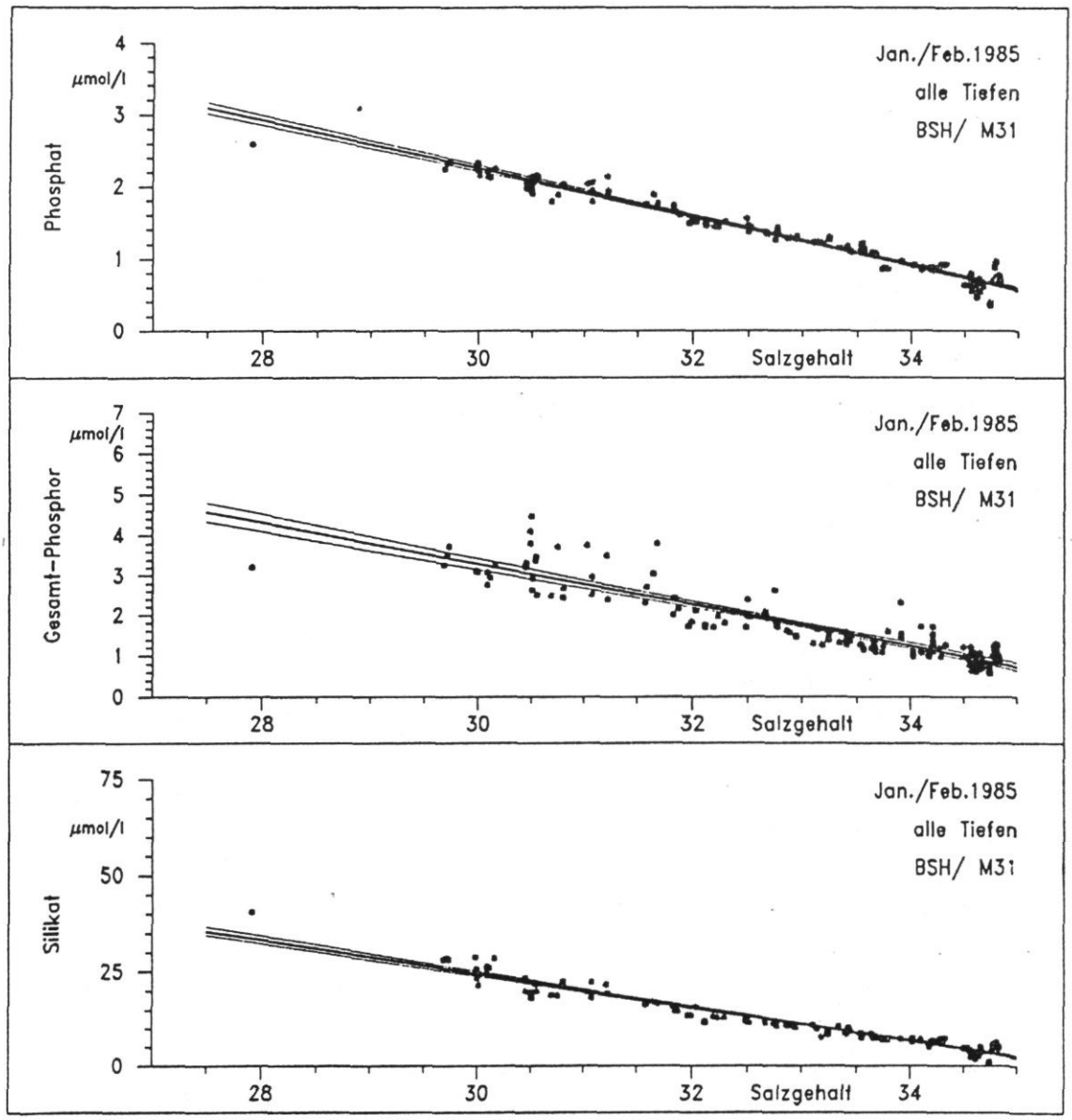

Figure 4a. Regression lines and 95% confidence ranges of the nutrient distribution in the German Bight in January/February 1985.

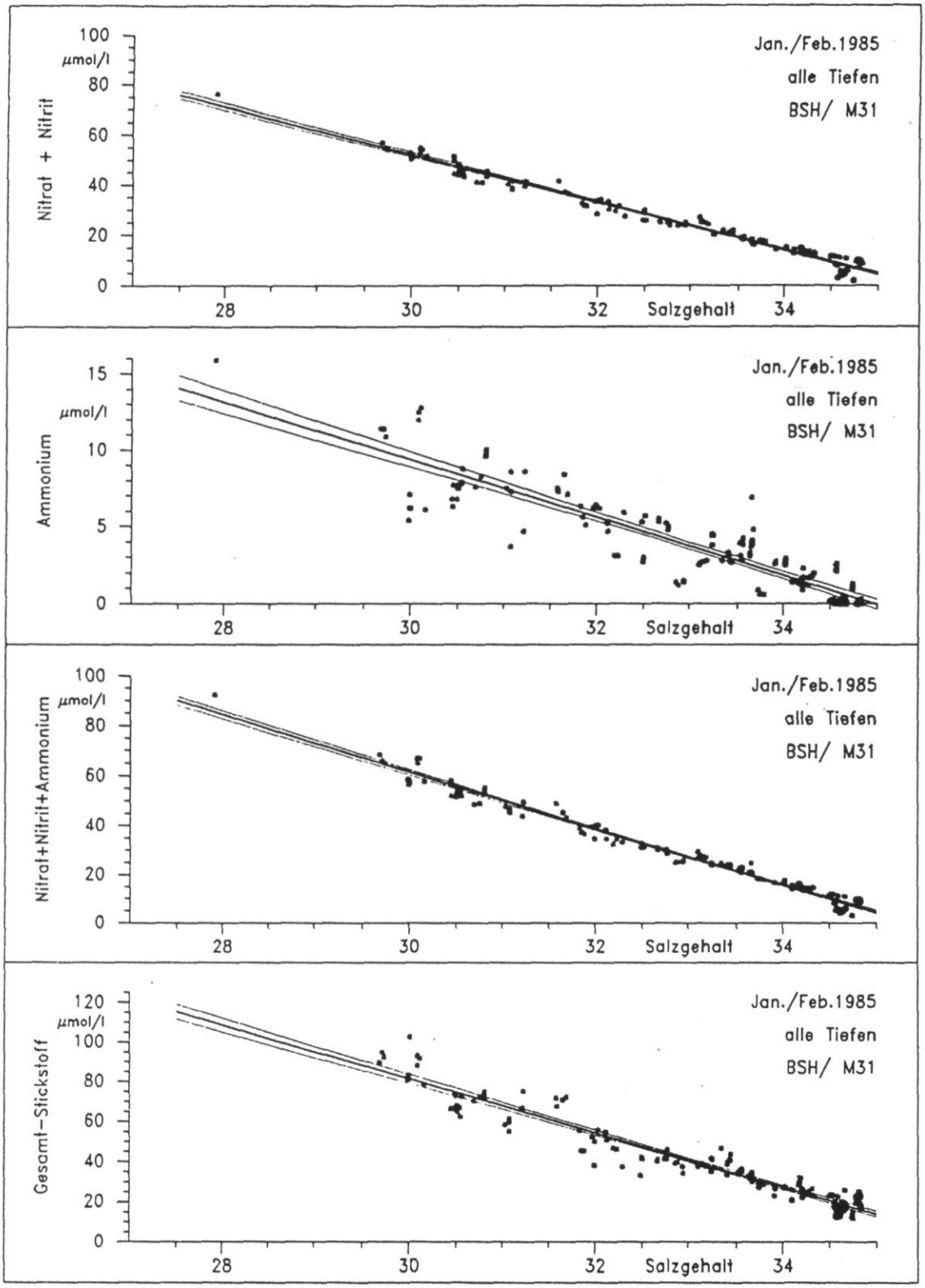

Figure 4b. Regression lines and 95% confidence ranges of the nutrient distribution in the German Bight in January/February 1985.

cant difference in the regression parameters is found in 1985 for total phosphate: the mean concentration at a salinity of 30 is 13% lower.

The regression analysis of the nutrient concentrations measured in January/February 1936 is given in Figure 6 and the relevant regression parameters in Table 8. The very slight slope of the regression line in the case of phosphate shows that the amount of nutrients entering the German Bight via rivers then was still relatively small. As a consequence the correlation coefficient is also very low.

Trend analysis

The mean concentration values and the 95% confidence ranges arising from the linear regression for salinity values of 30 and 33 are plotted in Figures 7–8. The salinity value of 30 is characteristic for a more or less near-coastal area depending on the hydrographic situation; the value of 33 is characteristic for areas in the central German Bight.

The figures confirm Weichart's results from 1986: compared with 1936, the phosphate concentrations in

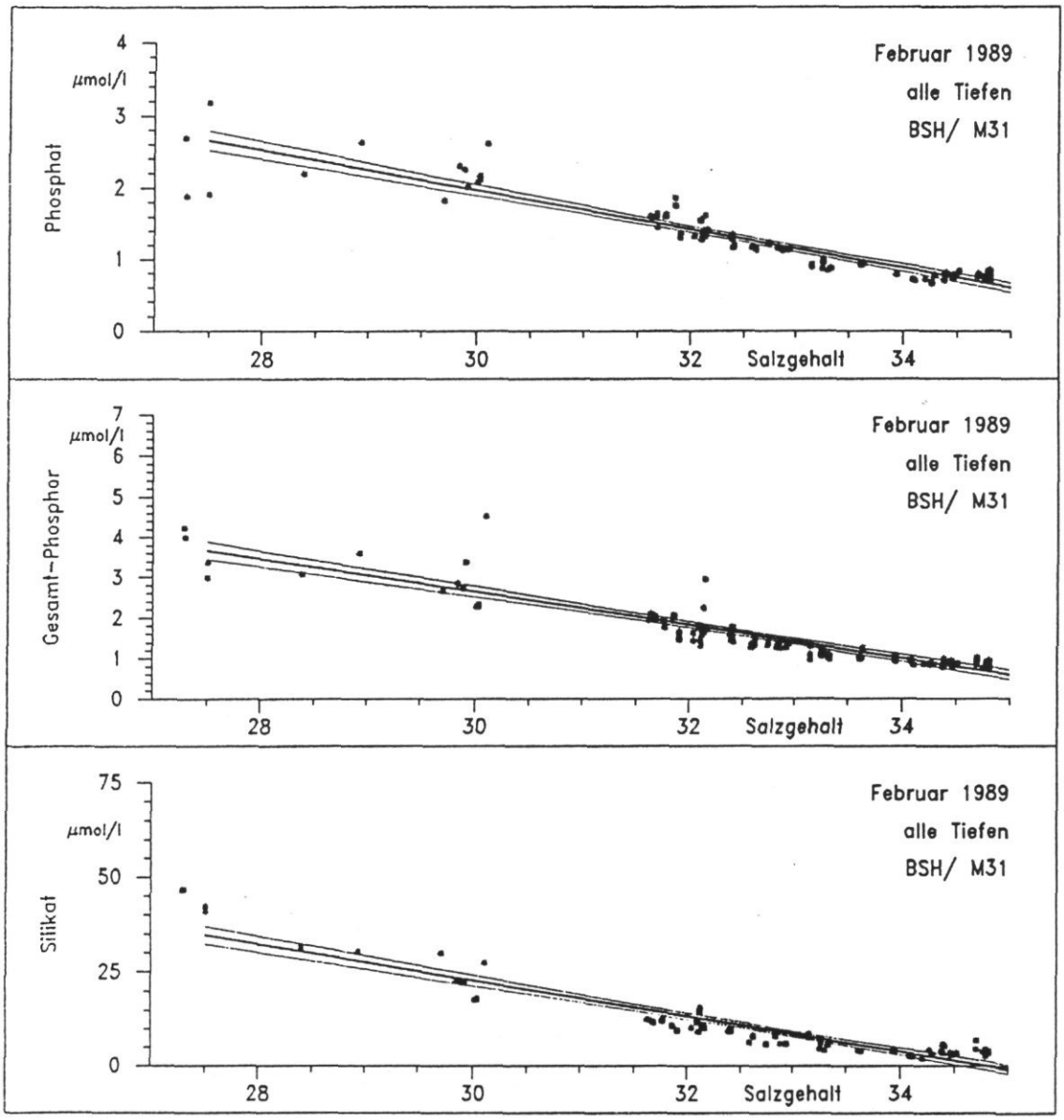

Figure 5a. Regression lines and 95% confidence ranges of the nutrient distribution in the German Bight in February 1989.

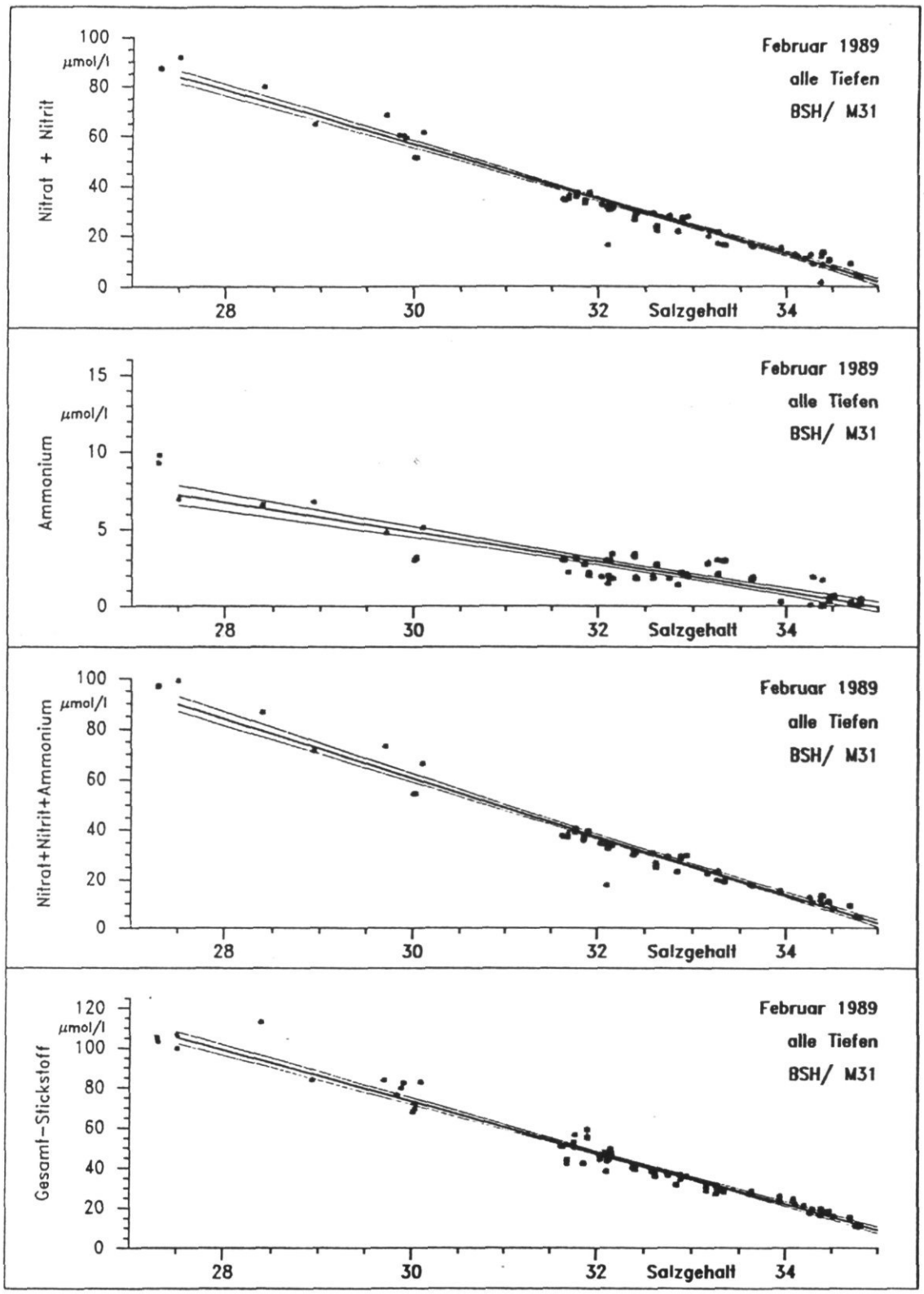

Figure 5b. Regression lines and 95% confidence ranges of the nutrient distribution in the German Bight in February 1989.

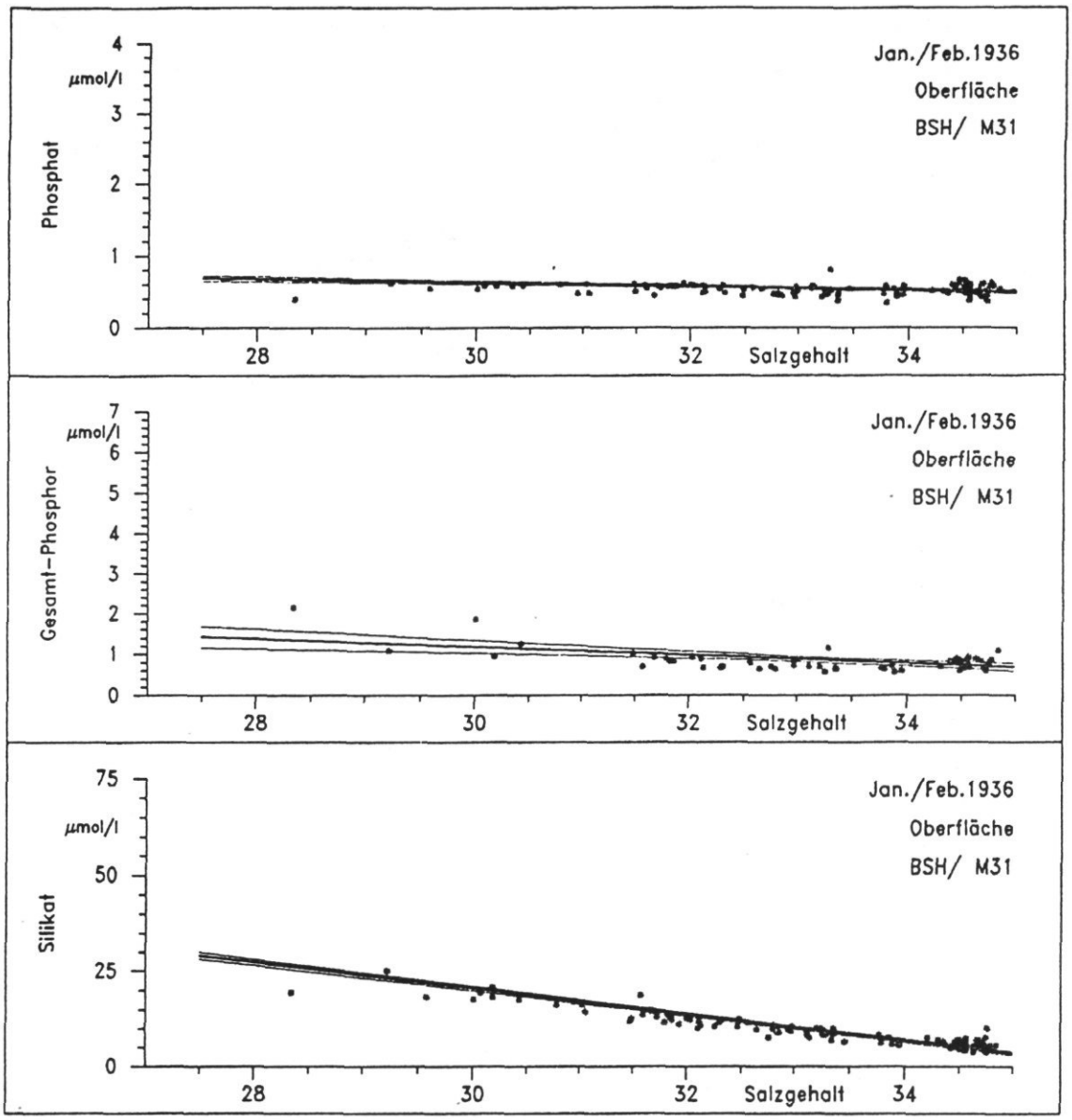

Figure 6. Regression lines and 95% confidence ranges of the nutrient distribution in the surface water of the German Bight in January/February 1936.

Table 8. Regression parameters of the nutrient–salinity relation in the surface waters of the German Bight in January/February 1936.

	No. of measurements	Slope	s.d.	Intercept	s.d.	Correl. coeff.
Phosphate	115	−0.03	0.00	1.39	0.09	−0.67
Total phosphorus	64	−0.10	0.02	4.17	0.58	−0.59
Silicate	117	−3.41	0.06	122.83	2.06	−0.98

near-coastal areas have greatly increased, while silicate concentrations have increased only a little.

Since 1978 (really, since 1985), no increase has been found in nitrogen compounds but, unfortunately, no comparison values from 1936 exist. The yearly fluctuations in the nitrogen compounds are greater than in other nutrients. A reason for this may be that nitrogen input can increase disproportionately when there is greater freshwater inflow (Körner, 1987).

Since 1985 the mean concentration values for dissolved phosphate and silicate have been somewhat lower than 1978. With the exception of the higher value in 1985, the mean concentrations of total phosphate are on the same level as 1978.

Our results from the stations near the island of Helgoland correspond very well with the data measured at Helgoland roads at the same time. From the nutrient data measured there each workday since 1962, long-term trends and annual concentration cycles were evaluated (Radach and Bohle-Carbonell, 1990). But as the island is situated in an area of frequently changing hydrographic conditions (Hickel and Berg, 1991), the data should be normalized to salinity, in order to evaluate short- and medium-term trends.

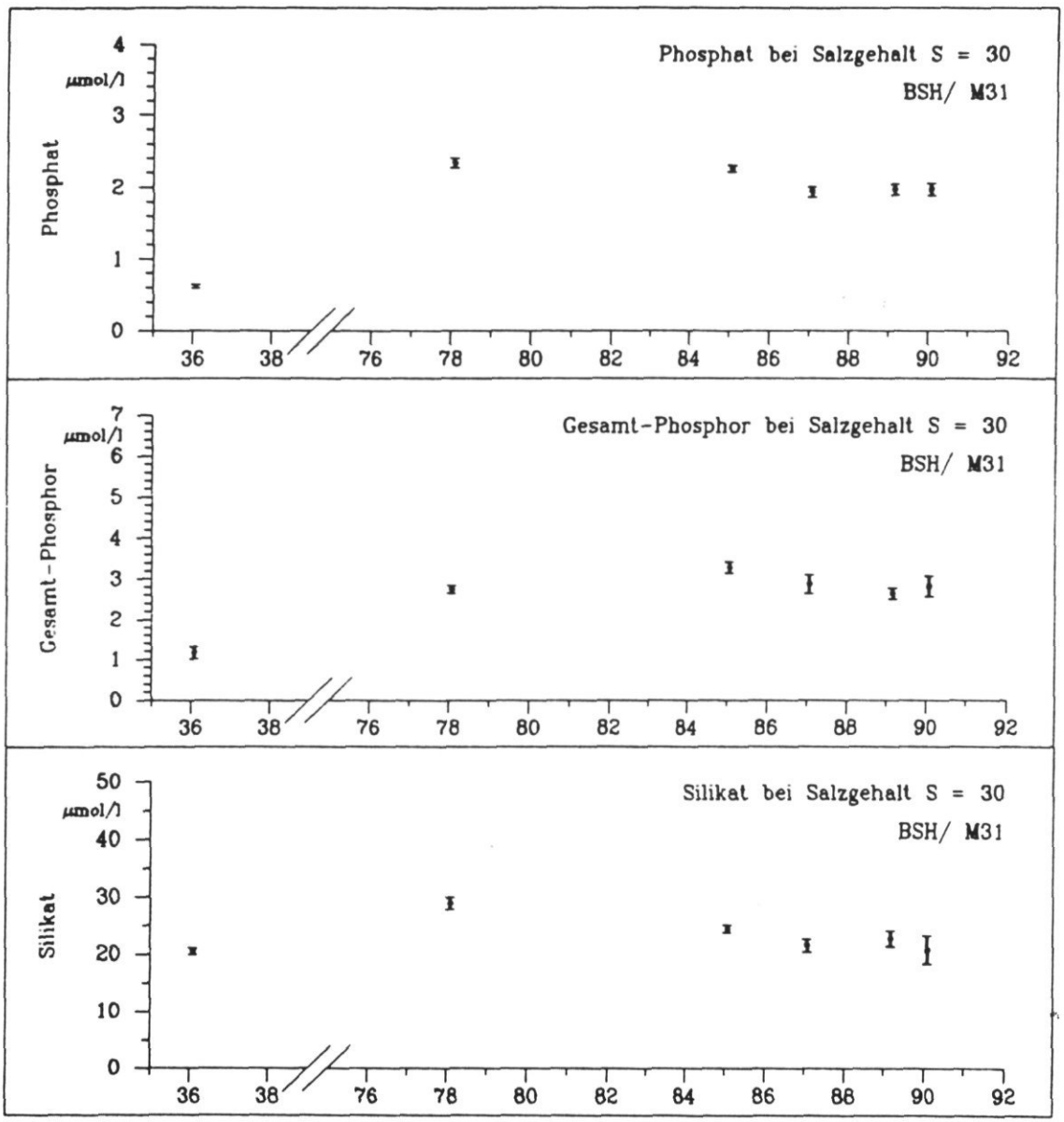

Figure 7a. Mean nutrient concentrations at a salinity of 30 in the winter water of the German Bight.

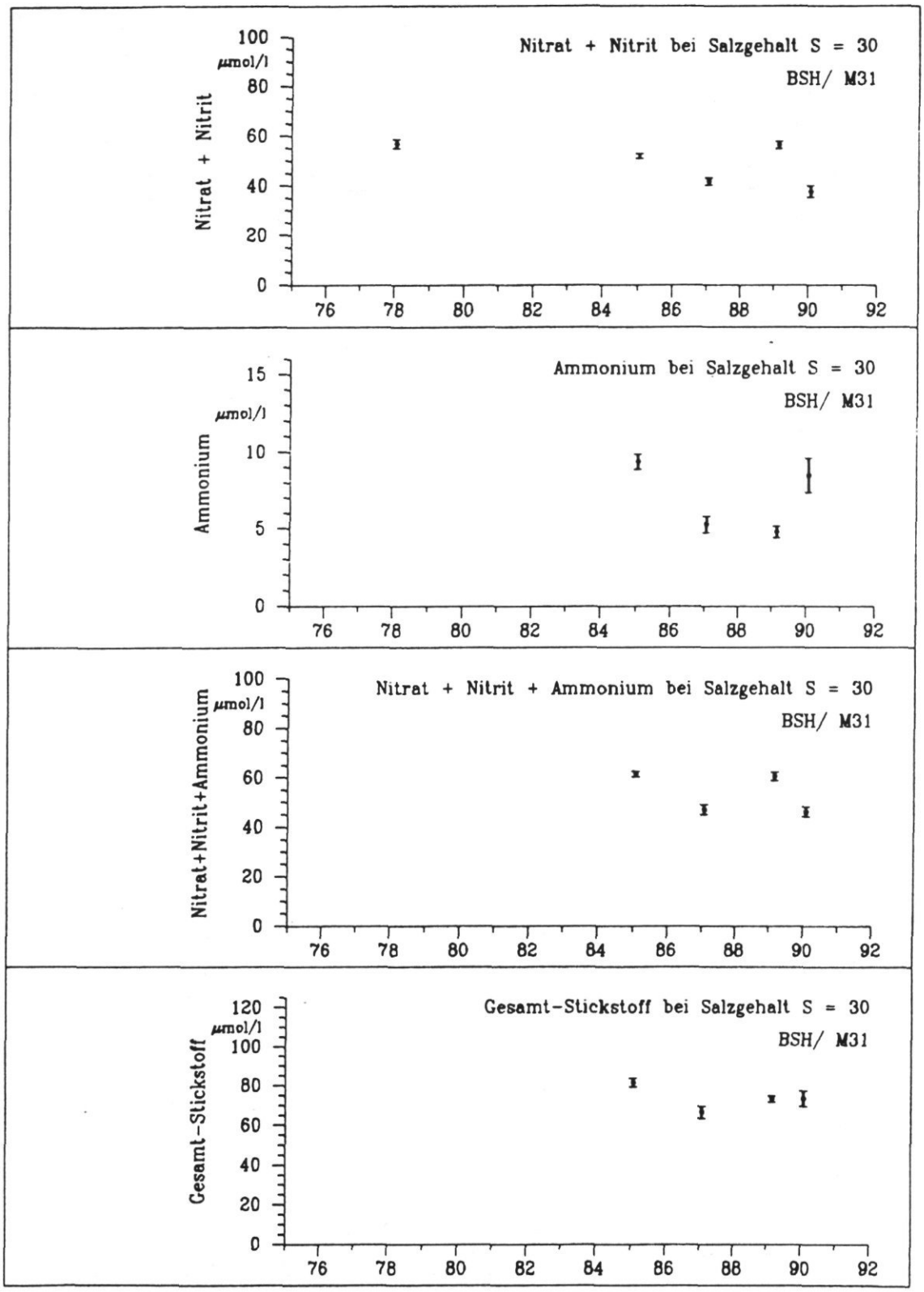

Figure 7b. Mean nutrient concentrations at a salinity of 30 in the winter water of the German Bight.

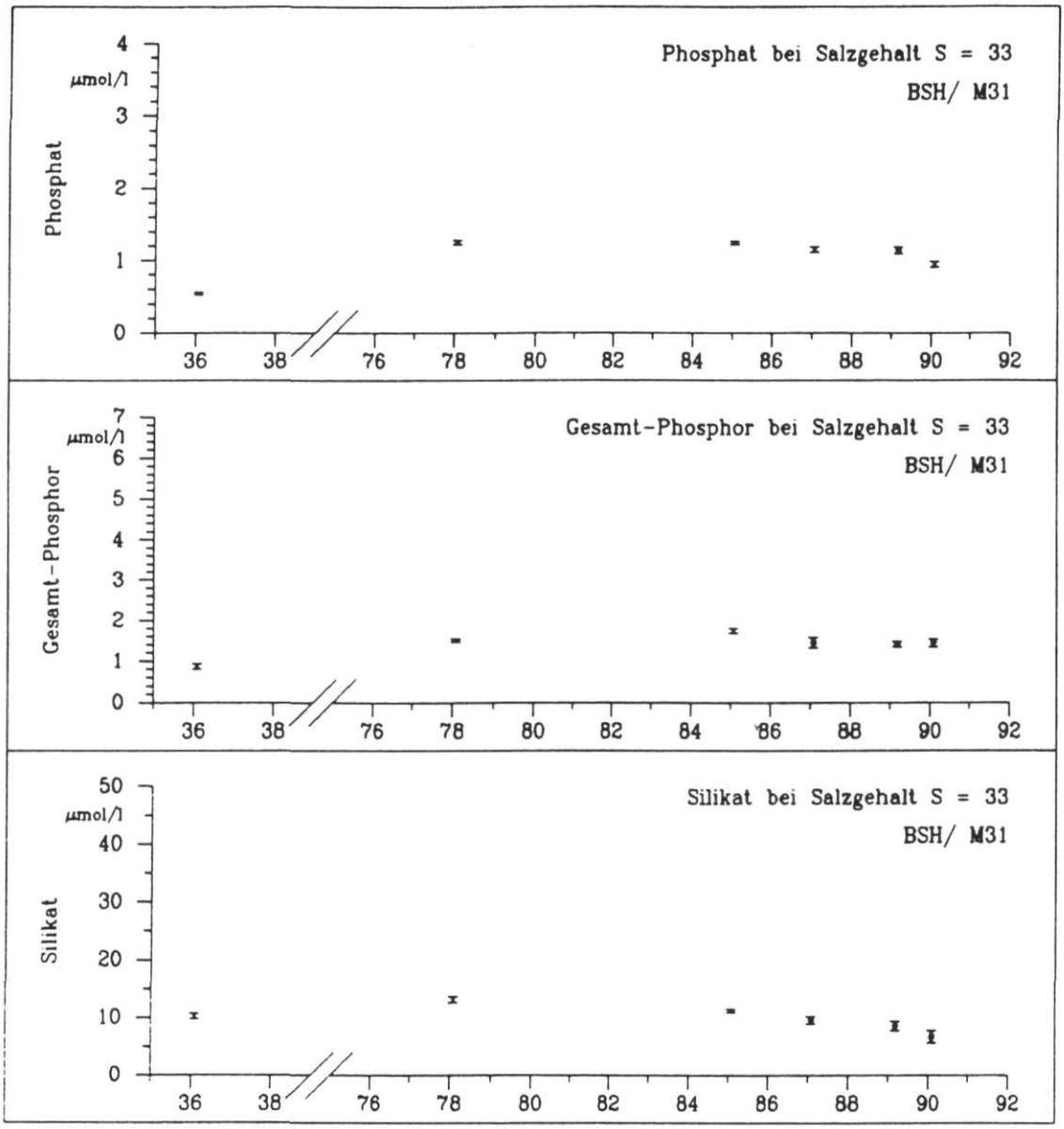

Figure 8a. Mean nutrient concentrations at a salinity of 33 in the winter water of the German Bight.

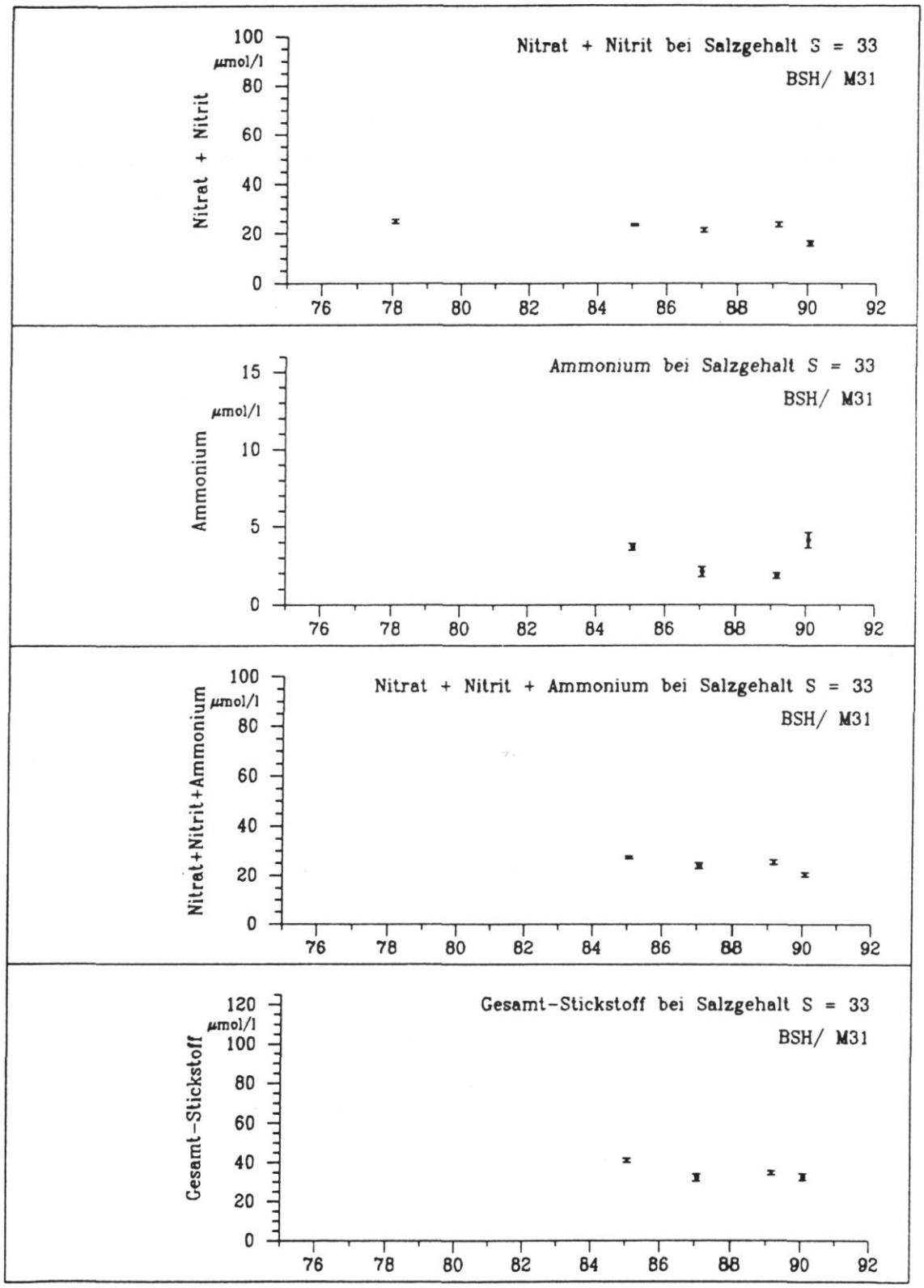

Figure 8b. Mean nutrient concentrations at a salinity of 33 in the winter water of the German Bight.

Nutrient winter data, measured in January–March 1987 in the entire North Sea, are published by Brockmann et al. (1990). In addition to our results obtained at the end of January 1987 in the German Bight, they give valuable information of the concentration distribution during winter in the other North Sea areas. The data confirm that the nutrient concentrations are highest during the winter months in the southeast of the North Sea.

Final remark: In January/February 1991 and 1992, the phosphate concentrations in the coastal water of the German Bight were 20% lower than in the years before.

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Sea-surface temperature: seasonal variation between the Iberian coast and the Madeira Islands, 1981–1987

Carlos M. Afonso Dias, Antonieta Amorim, and Maria das Dores Vacas

Dias, C. M. A., Amorim, A., and das Dores Vacas, M. 1992. Sea-surface temperature: seasonal variation between the Iberian coast and the Madeira Islands, 1981–1987. – ICES mar. Sci. Symp., 195: 177–186.

Sea-surface temperature (SST) data based on continuous thermographic records obtained on board two merchant vessels along the Funchal–Lisbon and Lisbon–Funchal routes in the period from January 1981 to December 1987 (331 records) are analysed. The SST mean seasonal variation obtained by harmonic analysis for 12 equally spaced points along the Funchal–Lisbon transect shows a maximum in August/September and a minimum in March with an average amplitude of 3°C. The influence of summer upwelling off the Portuguese continental coast is described. Year-to-year variations analysis reveals a general warming trend during the period 1981–1985.

Carlos M. Afonso Dias: Instituto Nacional de Investigação das Pescas, Av. Brasília, 1400–Lisbon, Portugal; Antonieta Amorim and Maria das Dores Vacas: Fisheries Research Laboratory-Biology Department 9000 Funchal, Madeira, Portugal.

Introduction

The seasonal variation of sea-surface temperature (SST) is the most pronounced and regular variation of any oceanic property. Its distribution and variation is caused by the combined effects of the heat exchange at the sea-surface, of dynamic advective factors, and of mixing by the wind.

In large parts of the oceans the annual global radiation is the principal factor that affects SST seasonal variation. Near the coasts, runoff and advection, particularly in coastal upwelling areas, can have a considerable influence in that variation. In consequence, the SST seasonal variation observed in most parts of the oceans behaves very regularly, approaching a truly sinusoidal variation.

Therefore, it is considered appropriate to describe the SST seasonal variation with a Fourier polynomial obtained by a least squares regression of digitized data collected on a monthly basis. Usually, truncated Fourier polynomials limited to the annual and semiannual terms are enough to describe significantly the SST seasonal variation (Wyrtki, 1965; Lynn, 1967; Panfilova, 1971).

Obviously, the harmonic parameters will not be directly suitable for a discussion of the SST seasonal variation causes. Their geographical distribution, however, can give an immediate view of the magnitude of the annual variation, and of its disturbance by the semi-annual variation, as well as provide information regarding its spatial coherence.

A description of the SST seasonal variation along an oceanic transect connecting the Portuguese continental coast (Lisbon) and the Isle of Madeira (Funchal) is presented in this paper.

Material and methods

Data set

The basic data utilized in this paper consists of SST continuously recorded (mercury-in-steel thermograph) at engine room intakes of two merchant vessels along the Lisbon–Funchal and Funchal–Lisbon routes during the period from January 1981 to December 1987 (Fig. 1).

Each voyage, nearly 510 nautical miles long, takes about 50 h following a direct course. The thermographs were checked against standard thermometers at both ends of the voyage; they were believed accurate to, at worst, about ±0.5°C. As a routine, the departure/arrival points (38°35'N–09°30'W and 32°35'N–16°50'W) were identified in the thermograph records by the ship's officers.

The number of observations (voyages) in each month of the period 1981–1987 is given in Table 1. The total number of observations was 331, with a lack of data from November 1985 to February 1986 and in July 1987. In about 80% of the 79 observed individual months there were three or more observations.

In order to make a digitized SST data set, twelve equally spaced points, including the extreme ones, were defined along the transect, as shown in Figure 1, and marked on the thermographic records. Their coordinates are shown under Figure 1.

The temperature readings on Points 1 to 12 for all individual observations (voyages) were assembled on a monthly basis (Fig. 1). The respective average values for each month of the period from January 1981 to December 1986 constitute the final data set.

Harmonic analysis

Considering data from each point as a time series, the SST mean seasonal variation was obtained by a least square regression of the data to a truncated Fourier polynomial of the form:

$$y = A_1 \cos\theta + B_1 \sin\theta + A_2 \cos 2\theta + B_2 \sin 2\theta + C \quad (1)$$

where y is the data value (monthly temperature) and is

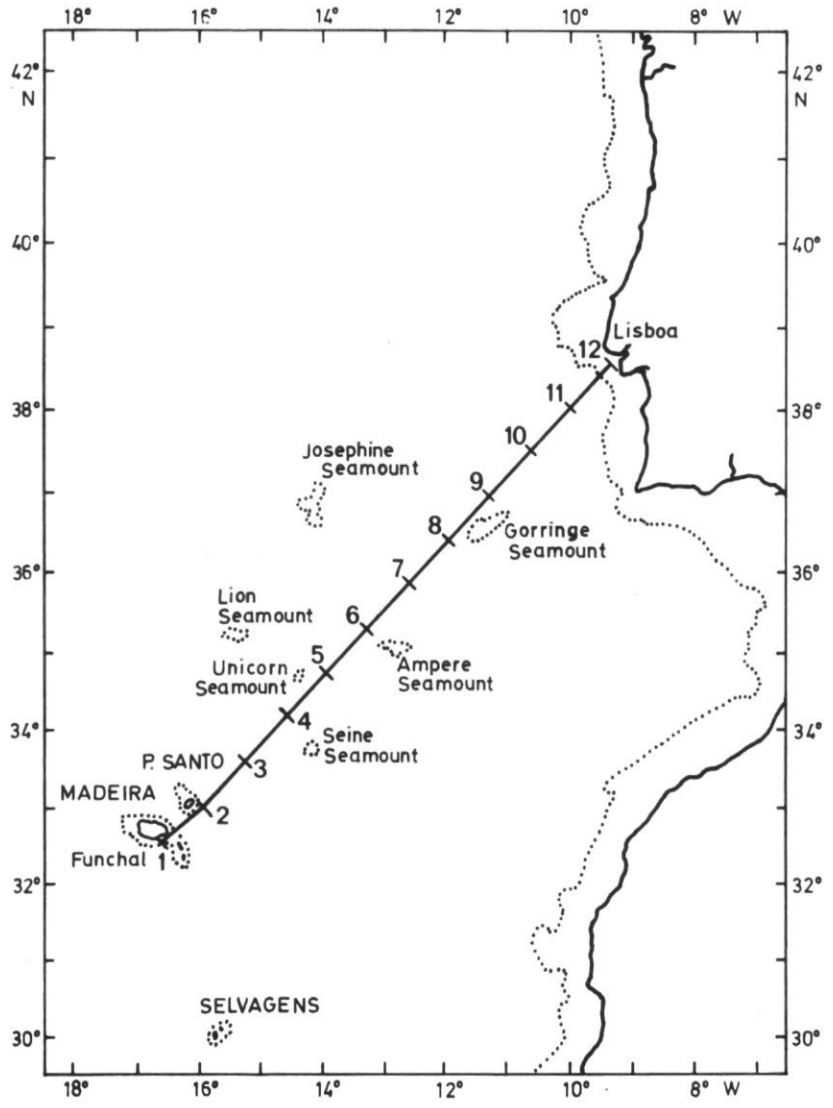


Figure 1. Location of the SST transect. The numbered points along the transect are the mean positions of the points where the continuous SST records were digitized.

Coordinates

P1	P2	P3	P4	P5	P6	P7	P8	P9	P10	P11	P12	
32°35′	33°08′	33°41′	34°13′	34°46′	35°19′	35°52′	36°24′	36°57′	37°30′	38°02′	38°35′	(N)
16°50′	16°10′	15°30′	14°50′	14°10′	13°30′	12°50′	12°10′	11°30′	10°50′	10°10′	09°30′	(W)

Table 1. Number of observations per month in each year.

	1981	1982	1983	1984	1985	1986	1987	Total
Jan	4	2	8	5	2	–	3	24
Feb	1	1	8	4	4	–	4	22
Mar	4	3	5	5	5	2	4	28
Apr	4	2	8	4	4	4	4	30
May	3	8	9	4	6	4	3	37
Jun	1	8	8	4	3	3	3	30
Jul	3	5	10	5	3	2	–	28
Aug	2	5	8	5	4	4	3	31
Sep	4	4	5	3	3	1	4	24
Oct	5	6	5	4	2	4	5	31
Nov	2	5	4	4	–	3	3	21
Dec	2	9	4	4	–	3	3	25
Year	35	58	82	51	36	30	39	331

the angular equivalent of the day of year (assumed at the middle of the respective month). For each point, the resulting set of 79 equations was solved simultaneously for the five coefficients using the Gauss–Jordan method.

The SST mean annual variation is given by the equation:

$$T(t) = T_0 + T_1 \cos(\alpha t - \phi_1) + T_2 \cos(2\alpha t - \phi_2) \quad (2)$$

where T(t) is the temperature for the time t (in days, starting from the beginning of January), at the phase angle (with $\alpha = 2\pi/365$), T_0 is the average annual temperature, T_1 and T_2 are the amplitudes, and ϕ_1 and ϕ_2 the phases of annual and semiannual waves. The coefficients of Equation (2) were obtained from those of Equation (1) following the usual procedures described in basic texts (e.g. Panofsky and Brier, 1968):

$T_0 = C$
$T_1 = (A_1^2 + B_1^2)^{0.5}$; $T_2 = (A_2^2 + B_2^2)^{0.5}$
$\phi_1 = \arctan(A_1/B_1)$; $\phi_2 = \arctan(A_2/B_2)$

Results

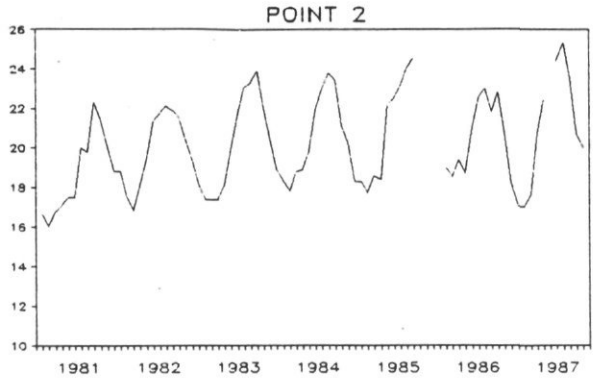
The SST monthly means observed at Points, 2, 4, 6, 8, 10 and 11 in the period from January 1981 through December 1986 (data set) are shown in Figure 2. Based on the same data the space–time representation of SST seasonal variation is presented in Figure 3.

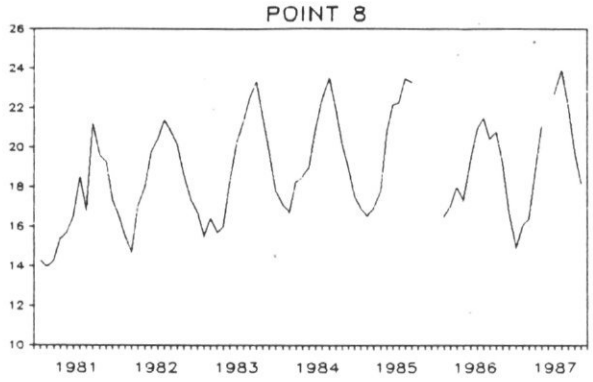
The coefficients of Equation (2) representing the SST mean seasonal variation regression curves are calculated and extreme values and range (temperature maximum minus temperature minimum) were derived. Statistics describing the fit of regression curves to data – standard deviation, multiple regression coefficient and estimate standard error – were also calculated. The results are given in Tables 2 and 3.

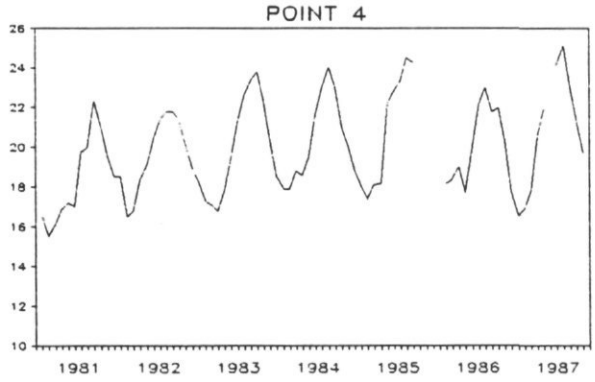
The long-term (1981–1987) monthly mean temperature obtained by digitizing the regression curves at mid-months (assuming 30 days for each month) are shown in Table 4. The related mean seasonal temperature variation along the Lisbon–Funchal transect is represented in Figure 4.

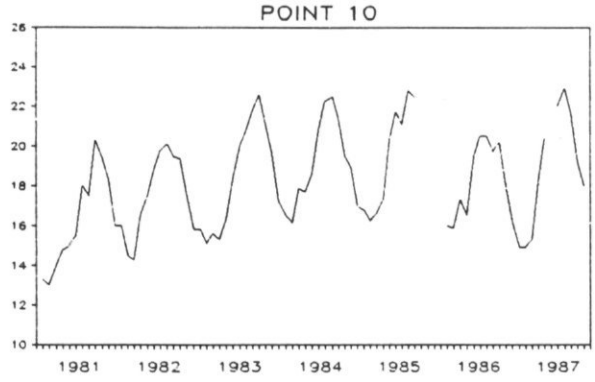
Table 2. Harmonic analysis. Amplitudes (°C) and phases (days) of the annual and semi-annual terms. $T(t) = T_0 + T_1 \cos(\alpha t - \phi_1) + T_2 \cos(2\alpha t - \phi_2)$ [$\alpha = 2\pi/365$]; R = coefficient of multiple regression (n=79); CD% = $R^2 \times 100$ = coefficient of determination.

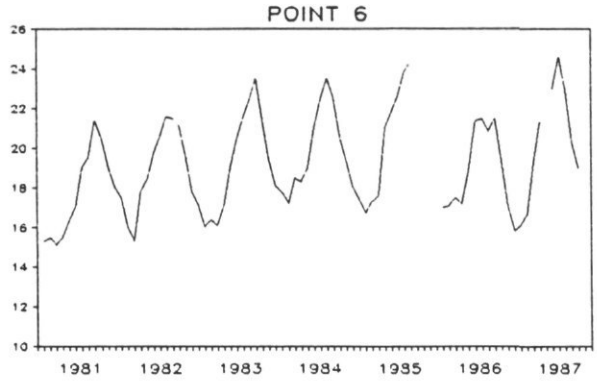
Points	T_0	T_1	ϕ_1	T_2	ϕ_2	R	CD%
1	19.99	2.87	249	0.27	214	0.917	84.1
2	20.12	2.89	250	0.35	213	0.897	80.5
3	20.05	2.92	250	0.38	207	0.894	80.0
4	19.88	2.99	251	0.42	202	0.899	80.8
5	19.59	3.00	252	0.43	200	0.903	81.6
6	19.15	3.10	252	0.44	205	0.905	82.0
7	18.98	3.18	249	0.50	211	0.899	80.9
8	18.78	3.18	249	0.42	215	0.892	79.5
9	18.47	3.16	249	0.32	222	0.888	78.9
10	18.15	3.10	247	0.37	220	0.872	76.1
11	17.60	2.91	246	0.30	216	0.844	71.3
12	16.47	2.41	237	0.18	358	0.786	61.8











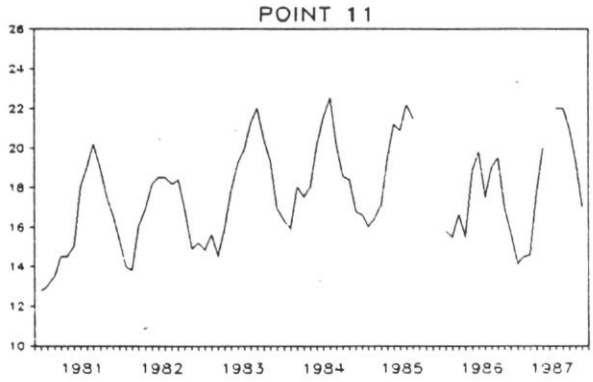


Figure 2a. Mean monthly SST variation during the period January 1981 to December 1987 at Points 2, 4, and 6 (see location in Figure 1).

Figure 2b. Mean monthly SST variation during the period January 1981 to December 1987 at Points 8, 10, and 11 (see location in Figure 1).

Figure 3 (opposite). SST seasonal variation along the Lisbon–Funchal transect during the period from January 1981 to December 1987. Temperature in °C.

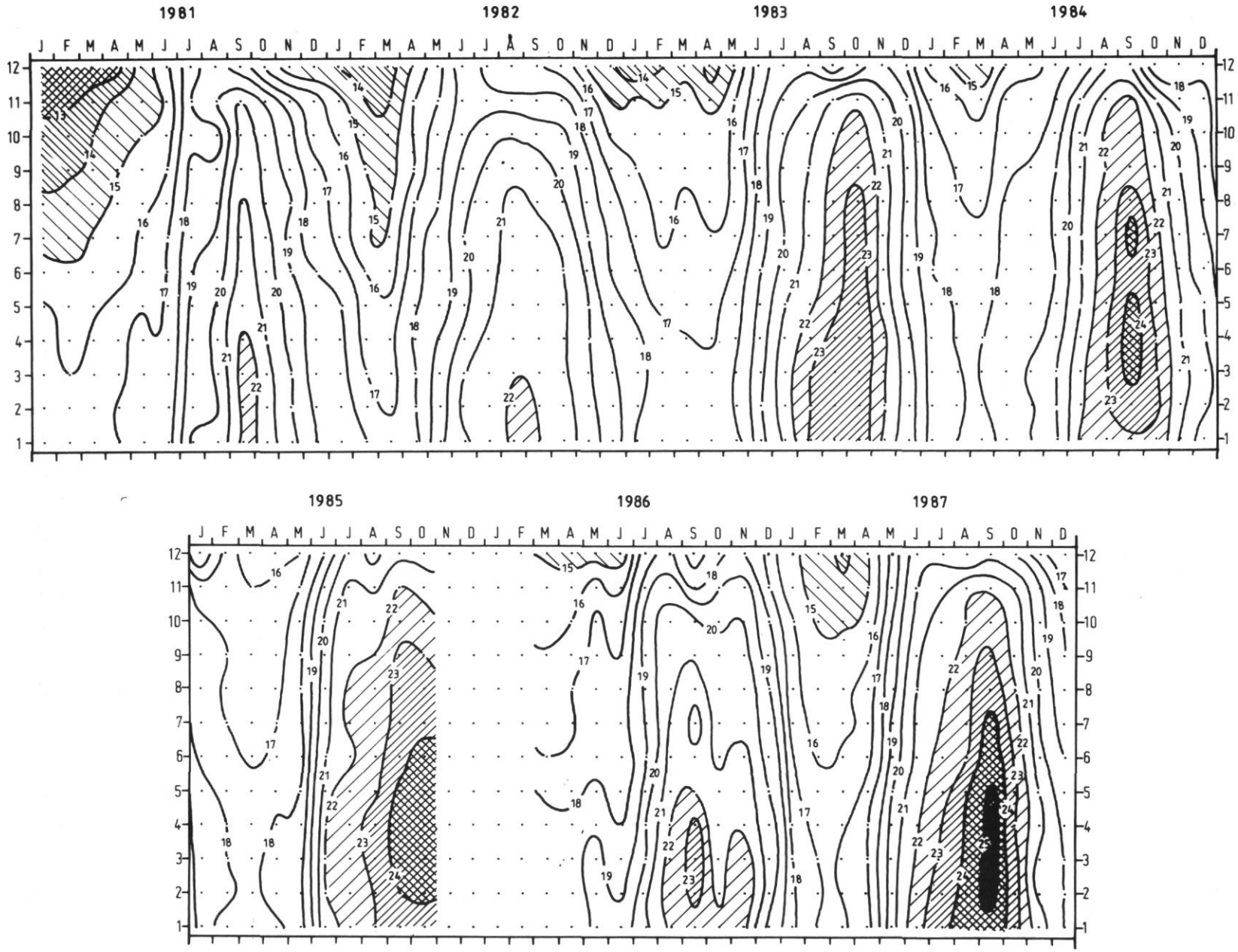

1981
1982
1983
1984
1985
1986
1987
J F M A M J J A S O N D

Table 3. Harmonic analysis. Temperature maxima (T_{max}) and minima (T_{min}), Ranges ($T_{max} - T_{min}$) and, in ( ), the respective dates (day of year). Values in °C. S = standard deviation about the 7-year mean. Se = standard error of estimate.

Points	T_{max}	(day)	T_{min}	(day)	Range	S	Se
1	23.0	(258)	17.1	(58)	5.8	2.245	0.920
2	23.2	(261)	17.2	(56)	5.9	2.338	1.052
3	23.2	(261)	17.2	(55)	6.0	2.354	1.081
4	23.2	(261)	17.0	(55)	6.2	2.403	1.083
5	22.9	(262)	16.7	(56)	6.2	2.398	1.058
6	22.5	(263)	16.2	(55)	6.4	2.471	1.077
7	22.4	(262)	15.8	(52)	6.6	2.551	1.145
8	22.1	(261)	15.6	(54)	6.5	2.574	1.191
9	21.7	(259)	15.3	(57)	6.5	2.556	1.207
10	21.4	(259)	15.0	(54)	6.4	2.553	1.280
11	20.6	(257)	14.7	(55)	6.0	2.467	1.358
12	18.8	(229)	14.0	(64)	4.9	2.193	1.397

Table 4. Mean monthly sea-surface temperature (°C) in Points 1 to 12 on the line Funchal–Lisbon in the period 1981–1987.

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)
Jan	18.0	18.1	18.0	17.8	17.5	17.0	16.6	16.5	16.3	15.9	15.5	14.8
Feb	17.2	17.3	17.2	17.0	16.8	16.2	15.8	15.6	15.4	15.0	14.7	14.1
Mar	17.3	17.4	17.4	17.2	16.9	16.3	16.1	15.8	15.4	15.2	14.8	14.0
Apr	17.9	18.1	18.0	17.8	17.5	17.0	16.9	16.6	16.2	16.1	15.6	14.7
May	19.0	19.1	19.0	18.7	18.4	17.9	17.9	17.7	17.4	17.2	16.7	16.0
Jun	20.2	20.2	20.1	19.8	19.5	19.1	19.1	19.0	18.7	18.4	17.9	17.4
Jul	21.4	21.4	21.4	21.1	20.8	20.4	20.3	20.2	20.0	19.7	19.1	18.4
Aug	22.5	22.5	22.5	22.4	22.1	21.7	21.6	21.5	21.1	20.8	20.1	18.8
Sep	23.0	23.2	23.2	23.1	22.9	22.5	22.4	22.1	21.7	21.3	20.6	18.6
Oct	22.6	22.8	22.8	22.8	22.5	22.2	22.0	21.8	21.3	21.0	20.2	17.9
Nov	21.3	21.5	21.5	21.4	21.1	20.7	20.5	20.3	20.0	19.6	18.8	17.0
Dec	19.6	19.7	19.6	19.4	19.2	18.7	18.4	18.3	18.0	17.6	17.0	15.9
Year	20.0	20.1	20.1	19.9	19.6	19.2	19.0	18.8	18.5	18.1	17.6	16.5

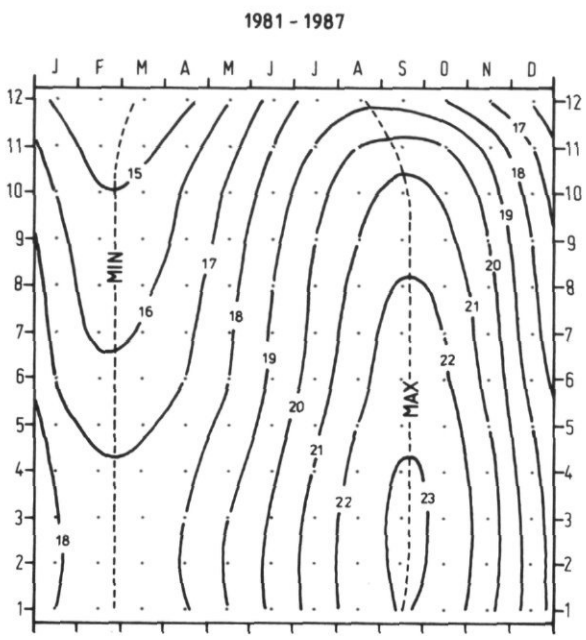


Figure 4. Mean seasonal SST variation along transect Lisbon–Funchal relative to the period 1981–1987. Temperature in °C.

Mean seasonal cycle

The SST mean seasonal cycle along the transect shows a normal north–south temperature gradient and a seasonal warming–cooling cycle within a range that varies from 4.9°C to 6.6°C (Table 3 and Fig. 5). Along all the transect the mean range increases from 5.8°C at Point 1 (Funchal) to 6.6°C at Point 7 (36.0°N). Between this Point and Point 10 (37.5°N), it is almost constant at around 6.5°C. This range significantly decreases with the proximity of the continental coast with the extreme value of 4.9°C at Point 12 (Lisbon).

The seasonal variation of the mean north–south gradient is only significant on the northern extreme of the transect (Fig. 6). Between Points 1 and 5 the mean monthly SST differences range from 0.1°C (September) to 0.7°C (June) and between Points 5 and 9 they range from 0.8°C (June and July) and 1.5°C (March). The seasonal differences in both cases do not seem significant. On the contrary, between Points 9 and 12 the gradient significantly increases, reaching a maximum in October with a mean monthly SST variation of 3.4°C.

The seasonal warming–cooling cycle is characterized by a minimum occurring along all the transect in

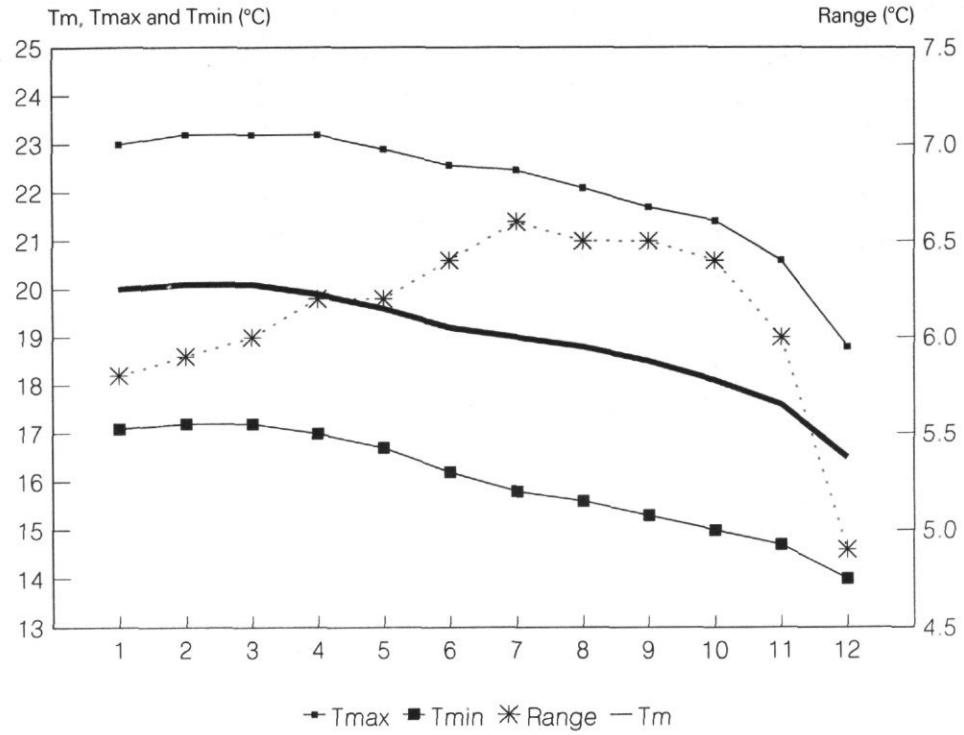


Figure 5. Mean, maximum, and minimum temperatures derived from the mean seasonal cycle at Points 1–12. ($T_{max} - T_{min}$) is also shown.

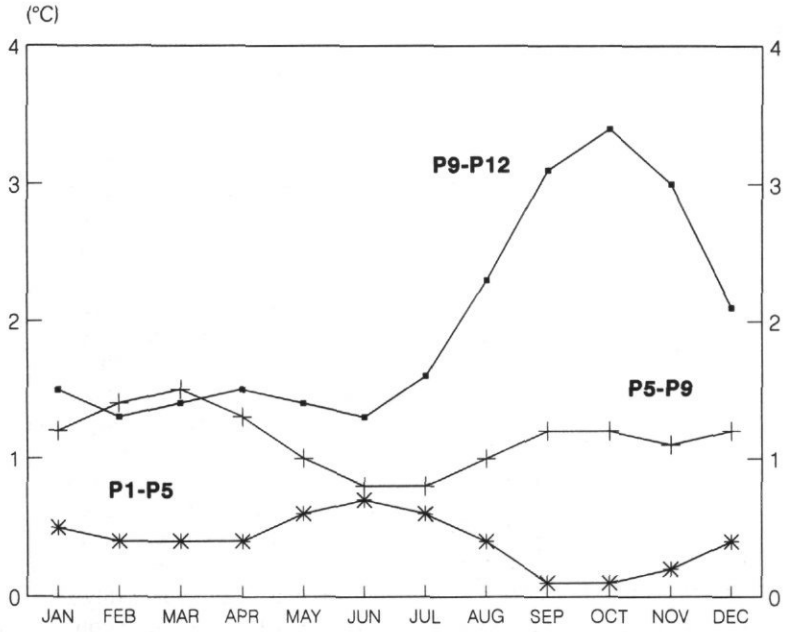


Figure 6. SST mean monthly differences between Points 1–5, 5–9, and 9–10.

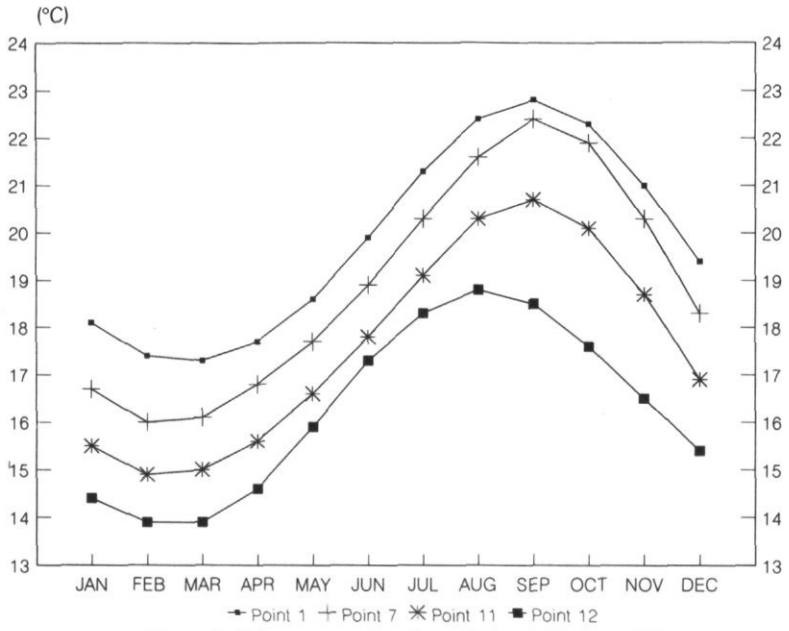


Figure 7. Mean seasonal cycles at Points 1, 7, 11, and 12.

February–March, and a maximum at all Points from 1 to 11 in September and at Point 12 in August (Table 3 and Figure 7).

Considering the latitudinal effect of the solar heating cycle, the seasonal cycles between Points 1 and 11 are similar. The differences between the seasonal cycles at Point 12 and those at other Points may be largely due to the effect of coastal upwelling. Significant effects in SST seasonal cycles can also be produced by differences between horizontal advection occurring in the coastal region and in the oceanic region and by spatial differences in ocean–atmosphere heat exchange (cloud cover).

Standard deviation about the 6-year mean (Table 3 and Fig. 8) distributes along the transect closely reflecting the mean range distribution. The coefficient of determination (square of the multiple regression coefficient) is a measure of the degree of relationship between the SST data and the time of the year:

$$\text{C.D.} = R^2 = \frac{\text{variance explained by the regression curve}}{\text{total variance}}$$

The distribution of this statistic (Table 2), which is inversely correlated with the standard error of estimate (Table 3), shows a moderate decrease from Point 1 (84.1%) to Point 10 (76.1%) and a more accentuated one from this Point to Point 12 (61.8%).

The low values of coefficient of determination at Points 11 and 12 can be explained by the effects of coastal upwelling. In fact, the sporadic nature of upwelling increases the total variance and the explained variance decreases when the effects of upwelling are out of phase with the solar heating–cooling cycle. According to the results of Wooster *et al.* (1976) and Fiúza (1982) on the Portuguese coastal upwelling variability and seasonality, a mixture of both effects is expected to occur.

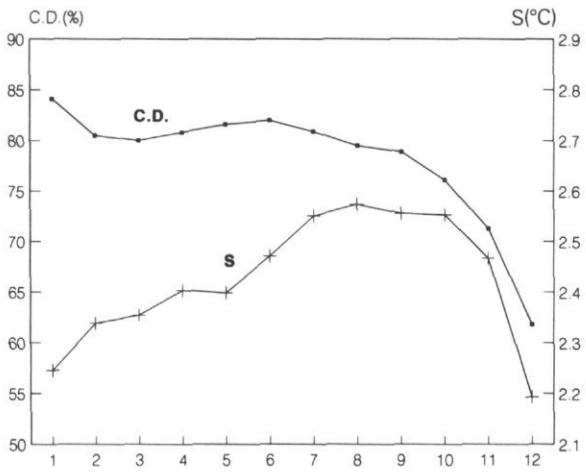


Figure 8. Standard deviations about six-year mean and coefficients of determination distribution along the transect.

Interannual variability

The SST seasonal variation along the Lisbon–Funchal transect observed during the seven-year period shows a considerable interannual variability. The most striking aspect of this variability is a warming trend all along the transect from 1981 through 1985. The mean warming rate during this period was about 0.6°C/year.

In order to visualize the warming trend along the transect, the monthly anomalies (SST observed minus the corresponding seven-year mean) were plotted for each one of the 12 Points. The resulting graphs relative to Points 2, 4, 6, 8, 10 and 11 are shown in Figure 9.

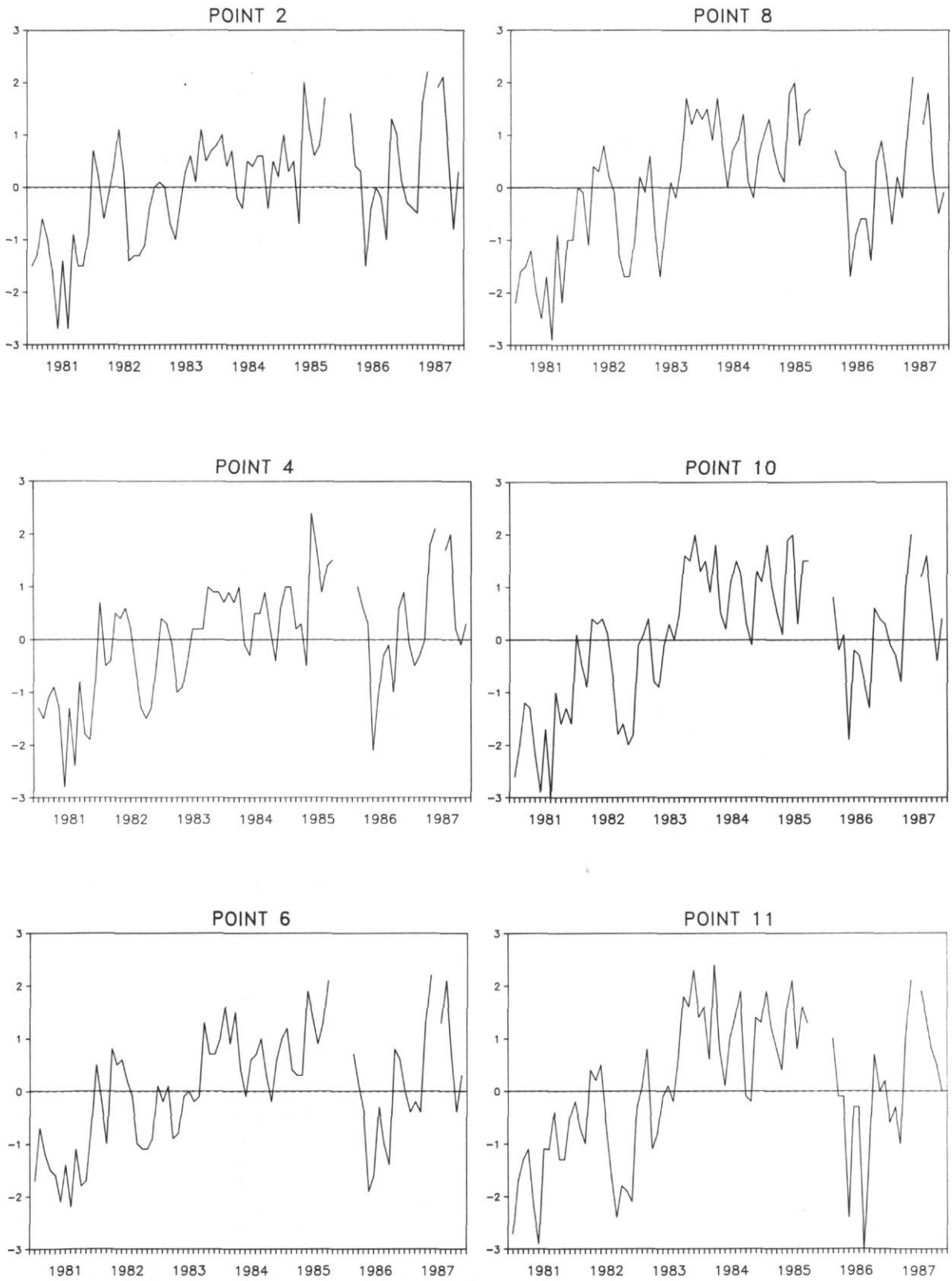


Figure 9. SST monthly anomalies referred to the 1981–1987 mean at Points 2, 4, and 6.

Conclusions

Recent publications describe important global climatic variability and anomalous warming events in the early 1980s (e.g. MacLain *et al.*, 1985; Horel *et al.*, 1986; Jones *et al.*, 1986).

In the Peru and California current systems extremely strong El Niño/Southern Oscillation (ENSO) warming events occurred in 1982–1983 (Quiroz, 1983; Reinecker and Mooers, 1986). In 1984 an anomalous warming event of the same type was observed in the Benguela current system (Shannon *et al.*, 1986). Philander (1986) showed that during the first half of 1984 oceanic and atmospheric conditions in the tropical Atlantic Ocean were similar to the conditions in the Pacific Ocean during El Niño. According to MacLain *et al.* (1985) in the Canary current system (15°N to 45°N) a general warming trend similar to that observed in the Pacific occurred during 1971–1984.

Despite the shortness of the time series and the poor quality of the data, the results presented here strongly suggest that in the region between the Portuguese continental coast and Madeira Island there occurred a progressive sea surface warming in 1981–1985.

This strong and fairly regular warming, followed by the significant cooling observed in 1986, seems to point out a transition between two extreme SST situations with a spatial scale large enough to include the Portuguese continental coast and Madeira Island.

A more conclusive description of the SST warm trend revealed here and a complete analysis and discussion of the large-scale atmospheric and oceanic processes involved in it are beyond the scope of this report. Nevertheless, the authors believe that the ecological importance of such an event, in particular for surface tunas, justifies further investigation.

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Mean sea level along the North Atlantic Spanish coast, 1980–1989

A. Lavín and M. J. García

Lavín, A., and García, M. J. 1992. Mean sea level along the North Atlantic Spanish coast, 1980–1989. – ICES mar. Sci. Symp., 195: 187–192.

The results are presented from an analysis of time series of sea levels and atmospheric pressures in three points of the North Atlantic Spanish Coast: Santander (43°28′N 3°48′W), Coruña (43°42′N 8°24′W), and Vigo (42°14′N 8°44′W). The period studied was from 1980 to 1989 in Santander and Vigo, and from 1980 to 1988 in Coruña. Time series of hourly sea levels were computed to get daily mean sea levels. A comparison is made between the power spectrum of the daily mean sea levels and atmospheric pressures, and the behaviour of their barometric factor. The highest value found of the barometric factor during 1989 in Vigo was −0.6 cm/mb. There is a significant trend of increasing values on the monthly mean sea levels in Santander and Vigo during the 1980s but not in Coruña. Highest variability appears in Vigo, especially during the first half of the decade. This place is situated open to North Atlantic influences and heavy seasonal upwelling occurs. In the second part of the decade all three points have similar behaviour. Seasonal effects are shown by higher values in autumn and winter and lower values during spring and summer. High variability occurs in Vigo when mean sea levels are 11 cm over the average level in November and 6 cm below it in summer. In Coruña and Santander those values are between 8 and 3.5 cm.

A. Lavín: Instituto Español de Oceanografía, PB 240 39080 Santander, Spain; M. J. García: Instituto Español de Oceanografía, Corazón de María 8,1 28001–Madrid, Spain.

Introduction

A useful signal of inter-year variability in ocean processes is provided by routine measurements of coastal sea level.

Sea level variations reflect contributions from a number of processes, including sea-surface slopes due to currents, effects of local winds, precipitation or runoff, atmospheric pressure variations and thermal effects (Montgomery 1938), and also contributions of non-locally forced sea height, which may propagate into the area as coastal trapped waves.

The response of sea to the variations in atmospheric pressure is not unlike an inverted barometer. The barometric factor does not always have a theoretical and constant value b = −1.01 cm/mb; the values are quite different and they are not constant. Robinson (1964) has given an explanation for this: high frequency atmospheric pressure variations (passage of frontal disturbances) can produce continental shelf waves due to resonance. Superficial winds may influence the mean sea level, but they may not be the only factor producing the non-barometric behaviour of sea level.

Material and methods

For this study, hourly sea levels were obtained from the Red Mareográfica of the Instituto Español de Oceanografía for three tidal stations situated on the North Atlantic Spanish Coast: Santander (43°42′N 3°48′W), Coruña (43°42′N 8°24′W), and Vigo (42°14′N 8°44′W) (see Fig. 1 for locations). They have been working since the 1940s. The period studied was from 1980 to 1989 in Santander and Vigo, and from 1980 to 1988 in Coruña.

A Xo Doodson filter was used to obtain the daily mean sea levels from the hourly heights, so that the tidal effects of a period of less than two days were eliminated. All values were referred to their own zero.

Atmospheric pressures for the decade have been taken from the Data Centre of the Instituto Nacional de Meteorología (Madrid) for three places: Santander, Coruña, and Marín. Daily values have been obtained as the average of the four synoptic values of the day.

Spectral analysis was done using the method of Blackman and Tukey (1958), with some modifications (Wagensberg and Estrada, 1978).

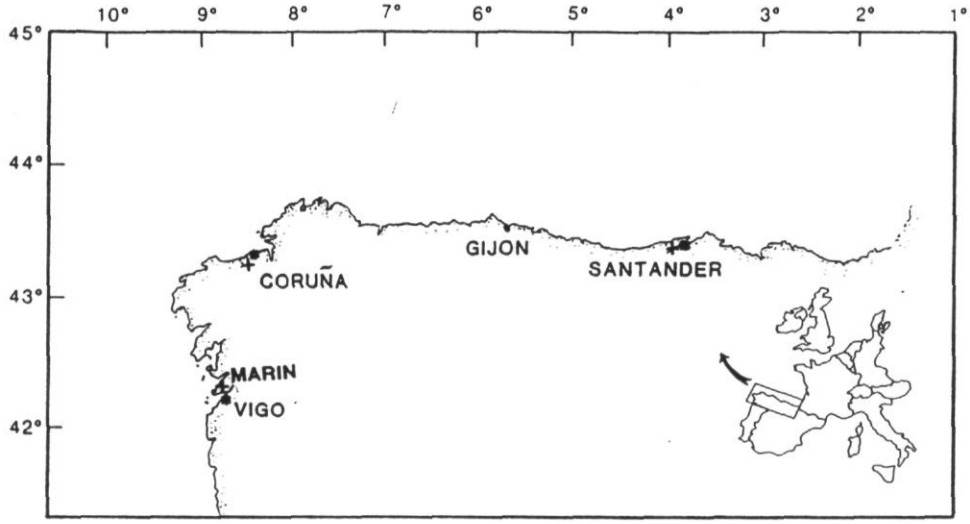

Figure 1. Location of tide gauge sites (*) and meteorological stations (+) along the northern Spanish coast.

Results

We show how variations of atmospheric pressure influence the daily mean sea level. In Coruña (Fig. 2a), sea level responds more like an inverted barometer (−1 cm/mb) than in Vigo (Fig. 2b). There must have been other meteorological factors, especially winds, that have more influence in Vigo than in Coruña.

The annual mean sea levels from 1980 to 1989 are presented for all three places (Fig. 3a) and are corrected by the effect of the atmospheric pressure fb = −1.01 mb/cm (Fig. 3b). All these values are anomalies referring to a mean value for the decade. There are no significant differences between the series. They show great variability during the first part of the decade, especially in Vigo where in 1981 the lowest value appears (−6.9 cm) and the highest in 1983 (8.3 cm) for the uncorrected data, and −7.1 cm and 9.6 cm for the corrected series (Fig. 3b).

The behaviour of the three series is similar for the second part of the decade, between 1985 and 1988, but differs in 1989.

We have made some regression equations (Table 1) to study the trend of the monthly means (for uncorrected and corrected data). A tendency to increasing values is found in Vigo and Santander, and changing values in Coruña, but the regression for this point is not significant. The highest tendency was found in Vigo with an increase of 0.64 cm/year.

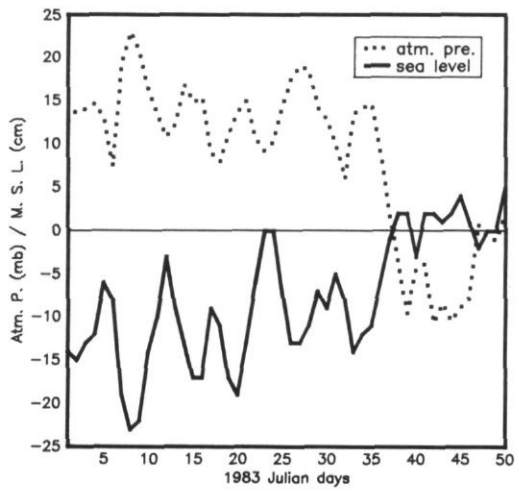

Figure 2a. Variations in atmospheric pressure and mean sea level in Coruña.

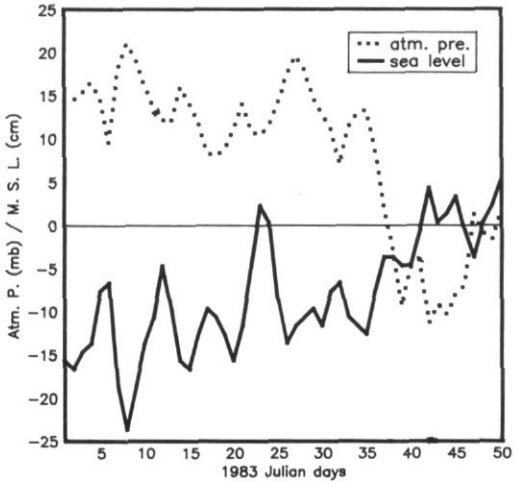

Figure 2b. Variations in atmospheric pressure in Marín and mean sea level in Vigo.

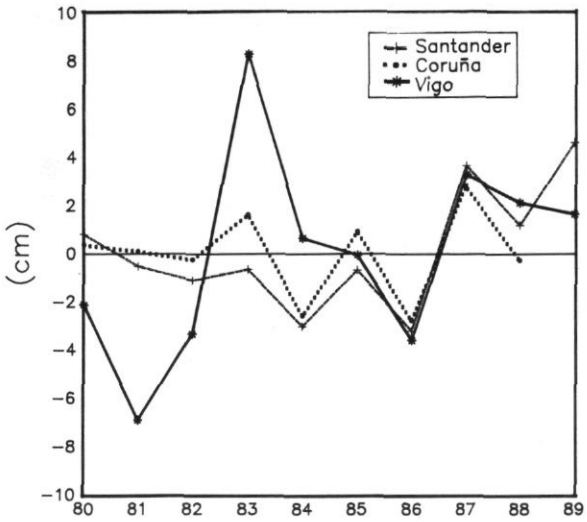

Figure 3a. Annual mean sea levels from 1980 to 1989.

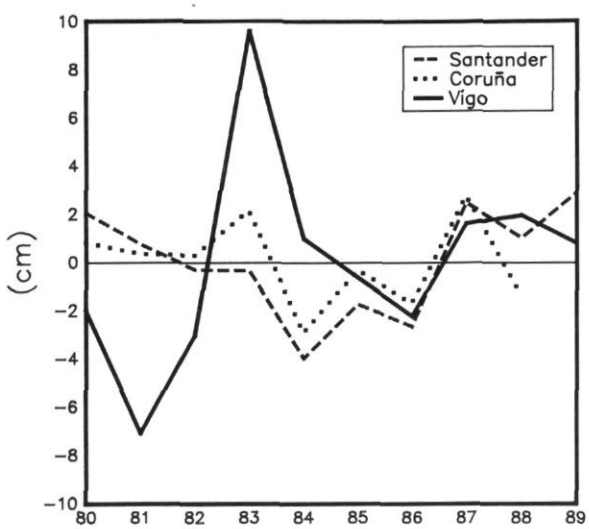

Figure 3b. Annual mean sea levels from 1980 to 1989 corrected by barometric factor.

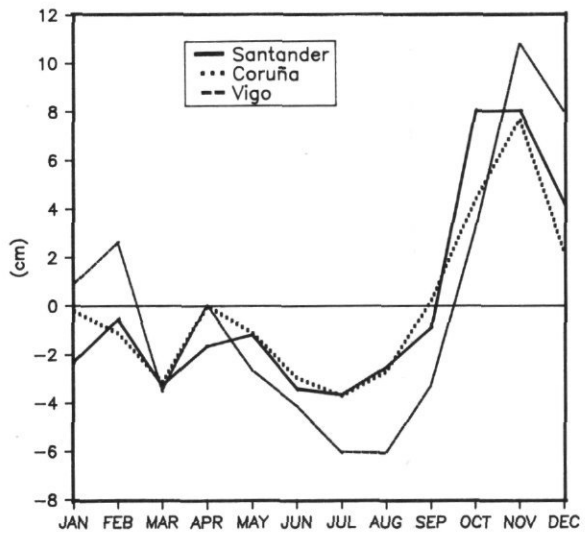

Figure 4. Seasonal mean sea levels for Santander, Coruña, and Vigo.

Table 1. Regression equations for mean sea level anomalies vs. time (months). Level of significance = 0.95 (* = ns).

	Interc.	Slope	n	r
Mean sea level Vigo	−3.23	0.053	120	0.20
Mean sea level Coruña	−0.41	0.007	108	0.04*
Mean sea level Santander	−2.44	0.040	119	0.20
Correct. m.s.l. Vigo	−2.54	0.042	120	0.17
Correct. m.s.l. Coruña	0.26	−0.005	108	0.02*
Correct. m.s.l. Santander	−1.21	0.020	119	0.12*

A comparison of slopes was made between the corrected and uncorrected values but no significant differences were found in either slopes or elevations in Vigo and Santander.

The mean of the monthly means for the three points referring to a mean value for the decade is presented (Fig. 4). They have a marked seasonality. Values are near average during the beginning of the year (January to April), lower during the upwelling season (May to September), and highest at the end of the year. There are also differences between Vigo and the other places. In Coruña and Santander the values ranged between −3.5 and 8 cm and in Vigo between −6 and 11 cm.

The upwelling index (Bakun, 1973) is calculated for the point 43°N 11°W (Lavín *et al.*, 1991) for 1981, 1983, and 1986 (Fig. 5). Considering the upwelling period (April–September), 1981 was an average year, 1983 had the lowest upwelling, and 1986 had the highest values of the decade.

Seasonal variations of the three sites for the mean sea level during the years 1981, 1983, and 1986 are presented in Figures 6a, b, and c. During those years we found very low levels in 1981, especially in Vigo, high values in 1983 (higher in Vigo), and in 1986 the values are a little under average for every month except February.

We observe in the power spectrum of atmospheric pressures in Coruña and Marín (Fig. 7a) two maximum values, at frequencies of 0.05 and 0.2 cpd (periods of 20 and 5 days). In Santander the maximum is found in a period of 10 days and the other values are similar to those of Coruña and Marín. These periods could be due to the passage of atmospheric fronts. These fronts arrived in Marín and Coruña directly from the Ocean, but have been distorted by land when arriving in Santander.

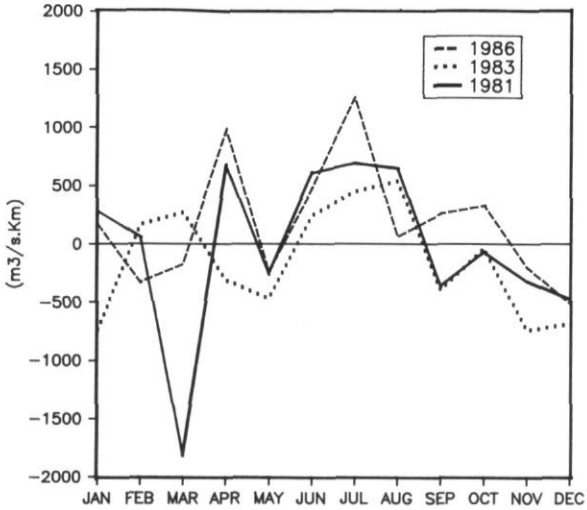

Figure 5. Upwelling index for the point 43°N 11°W during 1981, 1983, and 1986.

In the mean sea level power spectrum (Fig. 7b) of the three points we observe a first maximum (higher in Vigo) of 0.05 cpd and a second of 0.18 cpd. For shorter periods the energy is lower.

Comparing the power spectrum of mean sea level and atmospheric pressure for Coruña (Fig. 8a), we can see the coincidence of the maximum values as described earlier.

In Figure 8b the behaviour of the barometric factor as a function of frequency is presented. Similar behaviour is observed for Vigo and Coruña but values are higher in Coruña (ranging between 0.8 and 0.45) than Vigo (from 0.6 to 0.1). In Santander the values for the barometric factor range between 0.8 and 0.4, very close to those found for Coruña but not corresponding in frequencies.

In Table 2 we can see the correlations between data sets of daily mean sea levels, daily atmospheric pressures, northern and eastern components of wind, and corrected sea levels. Daily mean sea level appears highly correlated between Coruña and Vigo (0.84) and less between Vigo and Santander (0.66). Mean sea level and atmospheric pressure are more correlated in Coruña (−0.80), and Santander (−0.73) than in Vigo (−0.58). Wind effects are also negative with (−0.28) for the eastern component and (−0.35) for the northern one. For corrected sea level these values are a little lower, −0.20 and −0.33 respectively.

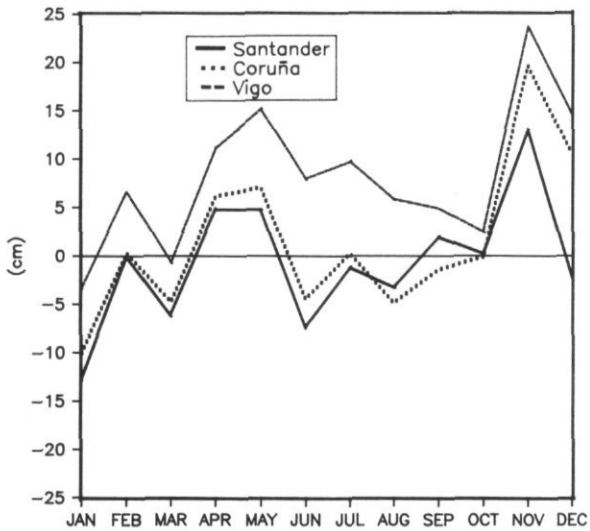

Figure 6b. Seasonal variations in sea level during 1983.

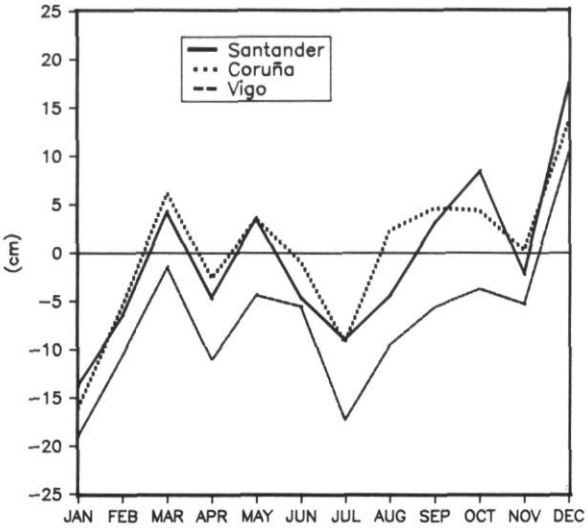

Figure 6a. Seasonal variations in sea level during 1981.

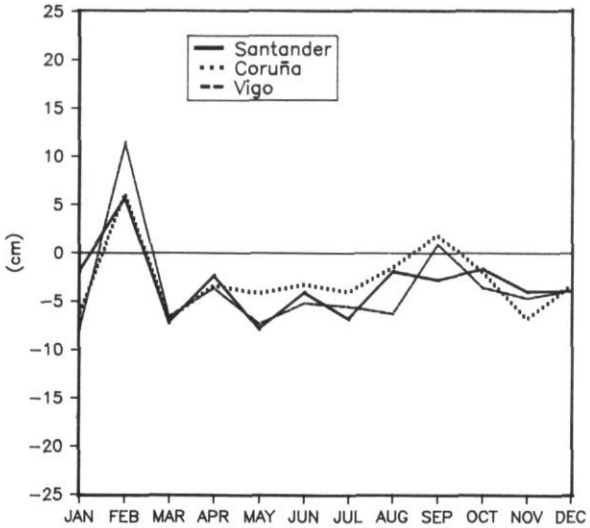

Figure 6c. Seasonal variations in sea level during 1986.

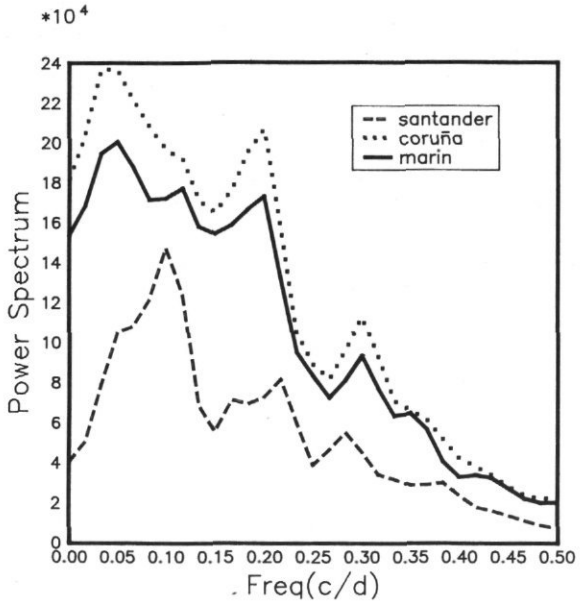

Figure 7a. Atmospheric pressure power spectrum.

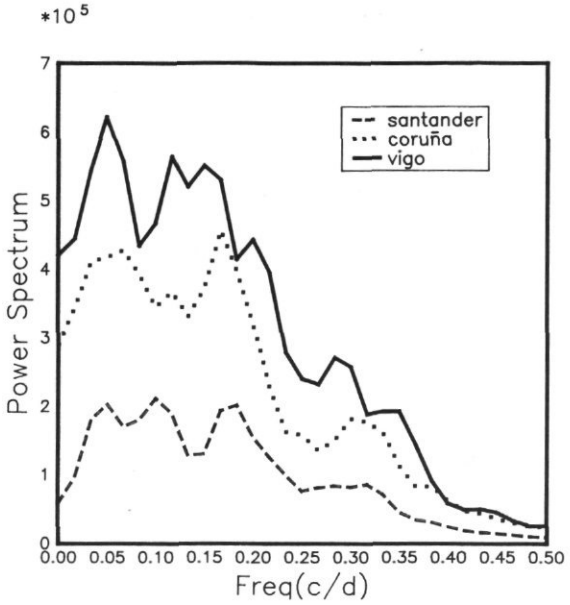

Figure 7b. Mean sea level power spectrum.

Correlations of wind components with atmospheric pressures are 0.20 for the East and 0.13 for the North. Finally, we have made the correlation between corrected mean sea levels and atmospheric pressures and we observe that it is very small.

Conclusions

We can observe different responses of mean sea levels in Santander, Coruña and Vigo to variations in atmospheric pressure. In Coruña and Santander sea level depends more on the atmospheric pressure than in Vigo. The barometric factor is higher. In Vigo the sea level is more greatly influenced by other meteorological factors.

We observe high variability in the first part of the decade, especially in Vigo with a maximum value for the mean sea level in 1983 and a minimum in 1981. There is a small trend of increasing values for the decade in Vigo and Santander but not in Coruña. There is a marked seasonality in monthly average levels with low levels in spring and summer and high levels in autumn. The power spectrums of the mean sea level and atmospheric pressure in Coruña and Vigo are similar but differ somewhat in Santander.

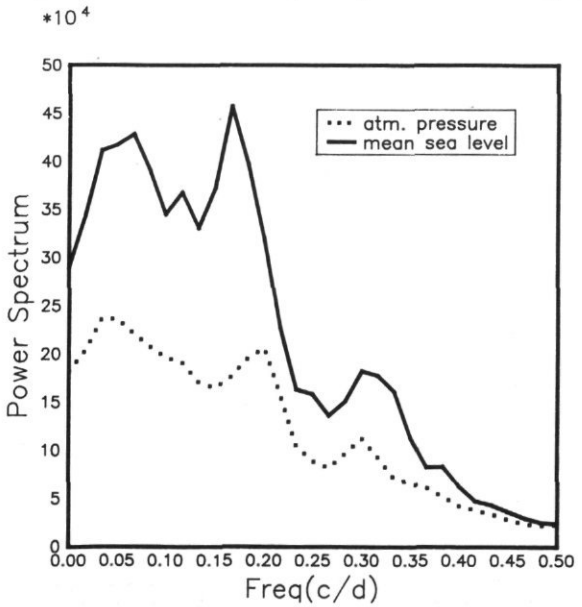

Figure 8a. Power spectrum of mean sea level and atmospheric pressure for Coruña.

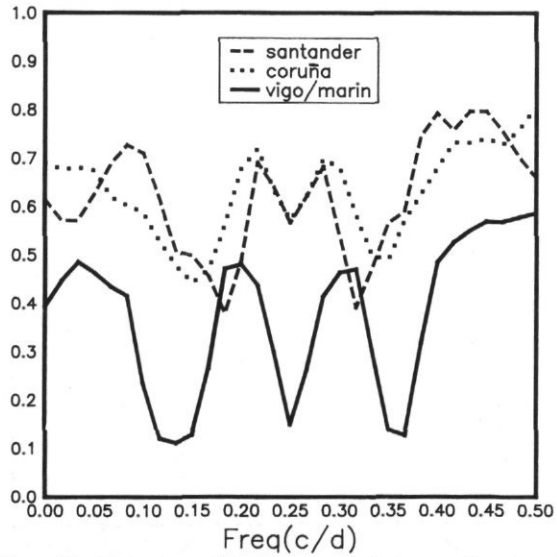

Figure 8b. Behaviour of the barometric factor as a function of frequency.

Table 2. Correlations between data sets of daily mean sea levels, daily atmospheric pressures, northern and eastern components of wind and corrected sea levels.

	M.s.l. Vigo	M.s.l. Coruña	M.s.l. Santander	At pressure Marín	At pressure Coruña	At pressure Santander	East wind Marín	North wind Marín	Cor. m.s.l. Vigo	Cor. m.s.l. Coruña	Cor. m.s.l. Santander
M.s.l. Vigo	*										
M.s.l. Coruña	0.84	*									
M.s.l. Santander	0.66	0.78	*								
At pressure Marín	−0.58			*							
At pressure Coruña		−0.80		0.96	*						
At pressure Santander			−0.73	0.90	0.96	*					
East wind Marín	−0.28			0.20			*				
North wind Marín	−0.35			0.13				*			
Cor. m.s.l. Vigo				0.01			−0.20	−0.33	*		
Cor. m.s.l. Coruña					−0.06					*	
Cor. m.s.l. Santander						0.07					*

M.s.l. = Mean sea level; Cor. m.s.l. = corrected mean sea level.

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Changes in plankton communities: when, where, and why?

Franciscus Colijn

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Through a growing interest and efforts, changes in plankton communities have recently been described and attributed to a whole series of causes ranging from climatological changes in the North Sea and Atlantic, ongoing eutrophication along coastal stretches of the North Sea and adjacent seas and bays, large infrastructural works, and changes in grazing pressure. The longest time series of plankton observations are derived from the CPR studies, whereas at a few locations (e.g., Plymouth, Texel, and Helgoland) time series are also available. Despite the wealth of information taken from the CPR the relevance of the data is restricted mainly owing to technical limitation of the CPR itself. The increasing observer effort combined with a rapid increase in ecophysiological plankton studies make it possible to better understand blooms of algae. This holds for well studied species such as *Phaeocystis*. The wax and wane of dinoflagellates is often attributed to specific hydrographic conditions, whereas the occurrence of species assemblages in summer is not understood owing to a lack of ecophysiological knowledge and the intricate interactions between trophic levels. The causes of succession are gradually being understood. Although eutrophication and its effects are among the most extensively studied phenomena they generally extend to a narrow stretch or areas of limited size. The effects can be temporarily devastating to other organisms. For most areas studied so far, however, effects are hardly distinguishable from natural variation. Understanding of the trend in natural events should, therefore, remain an important goal. Information from a Dutch monitoring study covering marine waters will be presented, together with the trends in nutrient concentrations and ratios. Algal species information will be updated.

Franciscus Colijn: Tidal Waters Division/RWS, PO Box 20907, 2500 EX The Hague, The Netherlands.

1. Introduction

In this paper the changes in phyto- and zooplankton in the ICES area (North Sea, Northeast Atlantic, and Baltic) over the period 1980–1990 will be described. All information is taken from existing literature, except for the Dutch coastal zone where new information has been compiled (Klein and van Buuren, 1992). Information on plankton communities in the ICES area stems primarily from studies with the Continuous Plankton Recorder (CPR). Surveys with this instrument have been carried out at monthly intervals since 1948 in the North Sea and the northeastern Atlantic (Lucas, 1973; Robinson, 1983). For the Baltic no such analyses are available and, therefore, one has to rely on a series of individual papers. During the last decade the analysis of CPR data has progressively been undertaken to investigate whether natural events such as climatic changes, or anthropogenic perturbations such as eutrophication, are the main causes for observed trends in plankton communities. In a few reviews, published recently, changes in plankton communities in the North Sea are treated (Reid *et al.*, 1990; Fransz *et al.*, 1991). Others (Smayda, 1989, 1990) have concentrated on specific effects of (toxic) blooms and their spreading on a global scale.

As a possible cause for an increase of blooms, both in frequency and intensity, eutrophication is very often mentioned (section 4). However, several examples can be given where hydrographic conditions or perturbations of the natural hydrography seem to be responsible for exceptional phytoplankton blooms (e.g. Holligan, 1987).

In this paper the results of the CPR studies will be presented first. Then the occurrence of phytoplankton

blooms in the ICES area will be discussed, based on all the observations described so far. Third, trends in species composition will be discussed for those sites where this information is available and, finally, an attempt will be made to analyse cause and effect using the results of recent (eco)physiological studies. In this way a link between hypotheses derived from field observations and experimental studies can be made.

2. Continuous plankton recorder data

Reid *et al.* (1990) have summarized some of the main facts regarding phytoplankton species based on CPR results for the North Sea, but most of their conclusions are related to changes already observed before 1980 (e.g. a trend of increased coloration of recorder silks in the southern North Sea until approximately 1975, and a decline thereafter). Analyses by Colebrook (1986) cover a part of the 1980s: because of a marked similarity both between species and between areas he concluded that environmental influences, such as the frequency of westerly weather, act on phytoplankton and zooplankton distribution and stocks (Fig. 1). In an earlier paper Colebrook *et al.* (1984) showed that a possible reversal of the downward trend in phytoplankton and zooplankton abundance had started around 1980. The variation on a monthly scale is very large, however, such that Zijlstra and de Wolf (1988) argued that it is hardly possible to discriminate between anthropogenic and natural influences. The data analysed by Colebrook (1986) also show that a general trend in the decline of zooplankton abundance is composed of simultaneous declines and increases of specific species. Some of the trends observed have been linked to sea-surface temperatures (Colebrook, 1985a), which would have an impact on overwintering zooplankton (Colebrook, 1985b). For phytoplankton, similar observations of species showing declining trends and those being almost

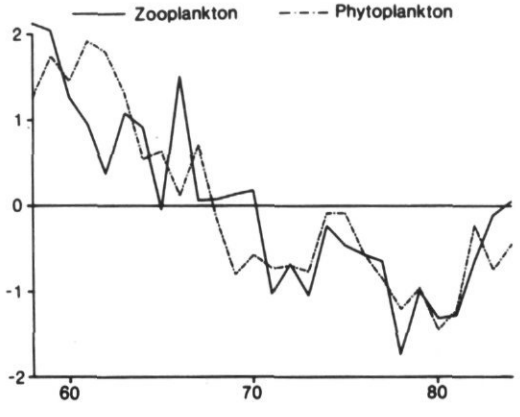

Figure 1. Trends in phytoplankton and zooplankton in the North Sea, based on CPR data (Colebrook, 1986).

stable in abundance with a strong variation have been observed (Robinson, 1983). Large-scale effects of climate on phytoplankton and zooplankton biomass have been studied by Dickson *et al.* (1988) and a peculiar example will be given by Dickson *et al.* (1992). They have concluded that a long-term increase in the northerly wind component over the eastern North Atlantic and the North Sea between 1950 and 1980 is associated with a decline in phytoplankton and zooplankton biomass. They assume that one of the causes for the phytoplankton decline could be attributed to a shortening of the growth period, which is dependent on the reduction of the mixed layer in the course of the growing season. This mechanism is based on Sverdrup's observations. Sverdrup stated that production starts whenever the mixing depth is less than the critical depth for growth. A shortening of the production cycle in the sea and a reduced rate of production thereafter would then be responsible for a decrease in food available for zooplankton. This hypothesis is supported by a comparison of two areas in the North Sea: a tendency for a diminution of the spring abundance peak and a shift in timing of the spring bloom from April to May were both observed. This observation was most clearly shown for the area where the maximum increase in storm frequency took place.

Another long-term trend was studied by Owens *et al.* (1989) (cf. Gieskes and Kraay, 1977). They examined long-term records of *Phaeocystis* obtained from the CPR. These have revealed that the occurrence of *Phaeocystis* has undergone a considerable decline over the study period (1946–1987) in the open sea areas of the Northeast Atlantic and the North Sea. A reversal of the trend is observed since 1980. The analysis also revealed a three-year cycle in the occurrence of *Phaeocystis*. Because of the coincidence of the trends in zooplankton and in the total phytoplankton community, the authors suggest the dominant influence of climatic factors.

A study by Reid *et al.* (1987) on blooms in the Northeastern Atlantic and North Sea based on CPR data will be referred to in the section on blooms (3.5). Trends in other areas, such as the Wadden Sea and the Dutch coastal zone, which are poorly sampled by the CPR, however, are caused by the predominant influence of eutrophication (Cadée, 1990a; section 3.1).

3. Trends in specific areas

In this section a few areas where intensive plankton studies have been performed will be compared and discussed: (a) the Dutch coast, including studies by Cadée (1992a, b and other references), Kat (1988), Klein and van Buuren (1992), and Koeman *et al.* (1991); (b) the English Channel, by Boalch (1987), and Maddock *et al.* (1989); (c) the German Bight and Helgoland, by Radach *et al.* (1986), Radach *et al.* (1990a), Radach and Bohle-Carbonell (1990), and Radach *et al.* (1990b),

and (d) the Baltic Sea, by Niemi (1982), Anonymous (1990a, b), and Smetacek *et al.* (1991).

3.1. Studies on phytoplankton in the Dutch coastal region

In 1989 a programme was started to follow and describe phytoplankton species composition in a quantitative way on four transects perpendicular to the coast (Fig. 2). The results of 1990 (Koeman *et al.*, 1991) will be presented here for the Noordwijk transect only. This transect has been sampled at 2, 10, 20, and 70 km respectively. The general species composition with the most obvious blooms (blooms have been defined for diatoms and flagellates including *Phaeocystis* when the cell numbers exceed 10^6, and for dinoflagellates when cell numbers exceed 10^3 per litre respectively) are given for stations N-2, N-10, N-20, and N-70 in Figure 3a–d, respectively. The hydrography of this area is characterized by a year-round mixed watermass which is under the direct influence of the river Rhine outflow at stations 2, 10, and 20, whereas station 70 has a watermass characterized by high salinity Channel water (see section 4.2.2). Station N-10 has been sampled almost weekly from April to September, the other stations with a biweekly frequency in spring and summer and monthly in autumn and winter. All the N-10 samples have also been counted and analysed with a flow cytometer before preservation. These data are compared below.

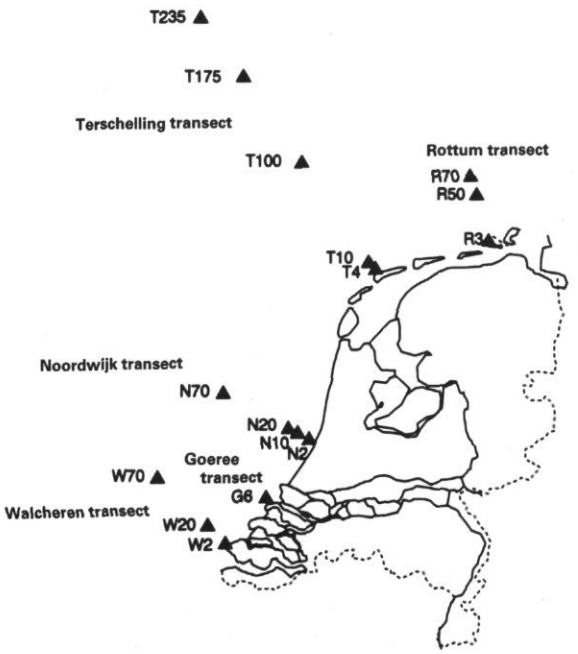

Figure 2. Transects and stations used in the Dutch biological monitoring programme (numbers refer to distances (km) from the coast).

At station N-2 (Fig. 3a) a spring diatom assemblage dominated by *Biddulphia aurita* var. *minima*, *Ditylum brightwellii*, *Melosira sulcata* and *Thalassiosira nitzschioides* was observed. At the end of April a dense *Phaeocystis* bloom occurs, which is succeeded by a period with high bloom levels of microflagellates in June and the first week of July. A particular species from the assemblage is *Rhodomonas*. At the end of July a bloom of *Mesodinium rubrum* accompanied by *Asterionella japonica* occurs, sometimes causing a red coloration of the water. Dinoflagellates, including species such as *Noctiluca, Gymnodinium,* and *Gyrodinium*, exceed the bloom level from July up to September. In early September a bloom consisting of different diatom species was observed.

Station N-10 (Fig. 3b) was sampled very extensively from mid-April to September. It showed two main periods of blooms: a conspicuous bloom of *Phaeocystis* in April, and a bloom of microflagellates, among others *Rhodomonas* in September, with in between, in August, a period of diatom blooms dominated by *Leptocylindrus*, *Chaetoceros*, and *Rhizosolenia* spp.

A comparable picture is shown by the results of station N-20 (Fig. 3c): the quantitative contribution, however, is different from the one shown for N-10. The *Phaeocystis* bloom is much denser with numbers exceeding 60×10^6, and diatom numbers much lower never reaching the bloom level of 10^6 cells per litre. The ratio of maximal numbers of *Phaeocystis* over diatoms, therefore, is about 85 in favour of *Phaeocystis* at N-20, whereas at N-10 this ratio was about 3. A possible explanation for this difference is the different silicate levels at N-10 and N-20, causing a much stronger silicate limitation at N-20. The last station on the transect, N-70 (Fig. 3d), is far offshore and not influenced by the eutrophic coastal watermasses. All cell numbers, therefore, are relatively low, with diatom numbers never reaching bloom level. *Phaeocystis*, however, does form a bloom in May, although it is delayed by two to three weeks as compared with the other stations on the transect. Central North Sea species observed at this station include *Prorocentrum micans* and occasionally *Gonyaulax* spp. A small diatom peak in June is dominated by *Rhizosolenia delicatula* and *Nitzschia* spp.

In conclusion, the phytoplankton species composition in the Dutch coastal zone is dominated in terms of numbers by flagellates. At times, however, diatoms take over in spring, but also during some summer months. Dinoflagellates in the coastal plankton occur in appreciable numbers only in June and July. For station N-10 the relative contributions of these three groups are given in Figure 4. When the diatom and dinoflagellate numbers are compared for all stations on an annual basis (Fig. 5) it is clear that diatoms are much more important in terms of numbers than dinoflagellates, except for station Ts-100, which showed stratification over a longer period

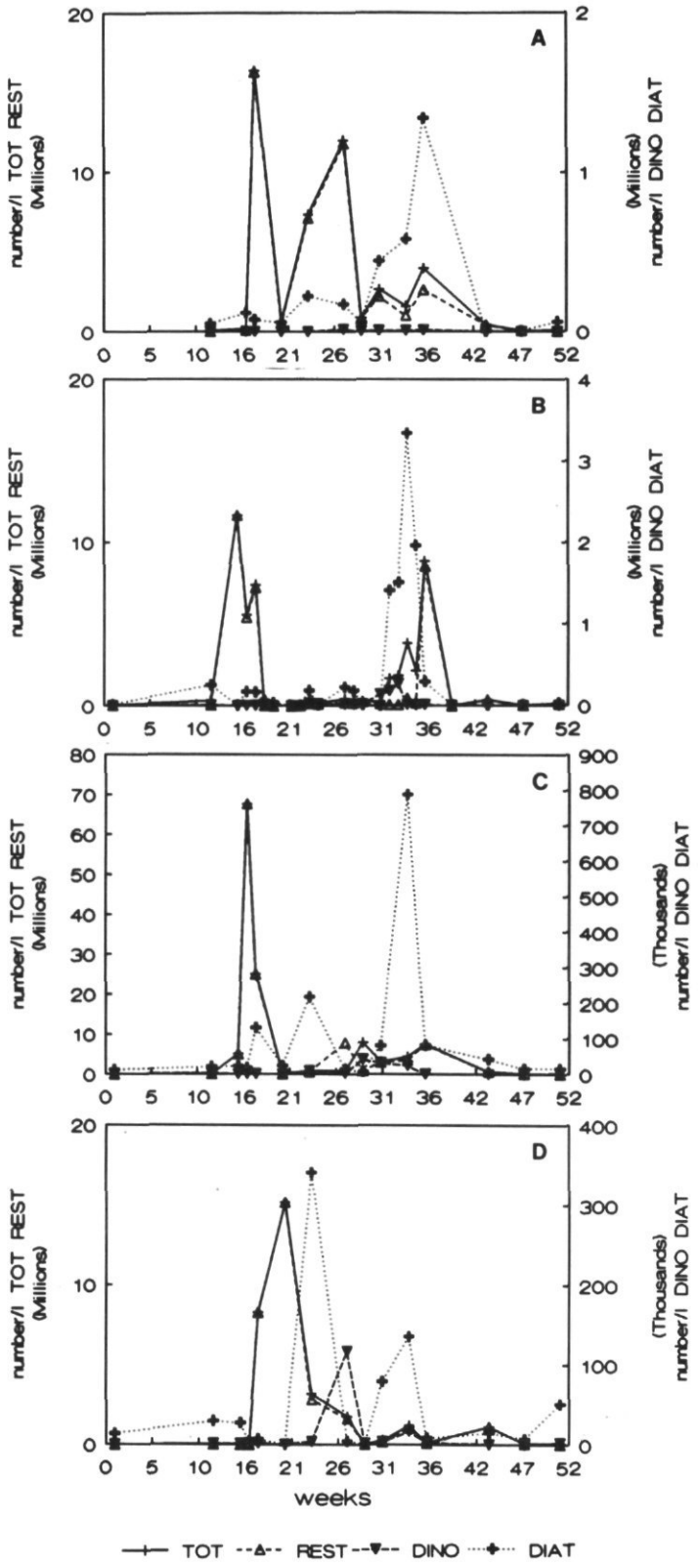

Figure 3. Phytoplankton succession in 1990 at station Noordwijk 2 (a), at station Noordwijk 10 (b), at station Noordwijk 20 (c), and at station Noordwijk 70 (d).

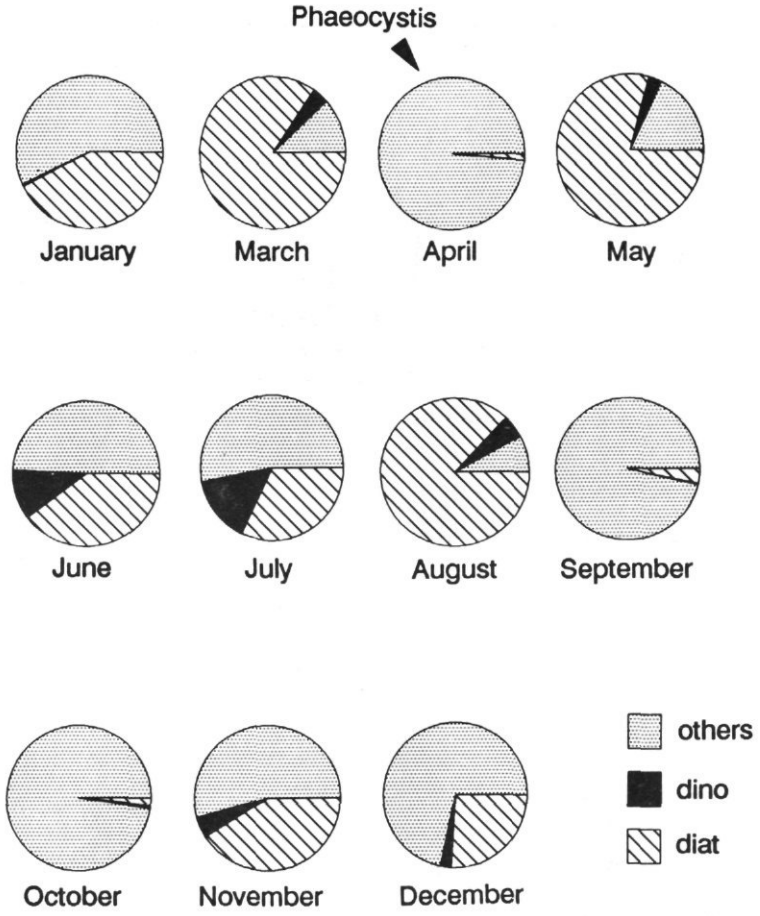

Figure 4. Average monthly ratio of diatom, dinoflagellate, and other species cell numbers at station Noordwijk 10 in 1990.

than any other station. The near coastal stations show a very pronounced dominance of diatoms.

Using the criteria stated, a total of 63 blooms of diatoms (3), *Phaeocystis* (22), *Rhodomonas* (3), and microflagellates (35) respectively, were observed in the Dutch coastal zone for the whole of 1990. However, taking into account sampling frequency and stations the number of *Phaeocystis* blooms reduces to one single bloom in mid-April on almost all stations and a second bloom observation in July near Terscheling (T4). The number of blooms of microflagellates can also be reduced to a late summer period extending from June to October. In this case the species composition may have changed considerably. Better criteria (e.g. based on biomass) are therefore needed to describe and compare the densities of blooms.

When we compare the results from this phytoplankton monitoring study with older data (Kat, 1988, Gieskes and Kraay, 1975) there is no reason to suggest long-term changes in species composition. A comparison with the data analysed by Kat (1988) is based on her Figure 4. She showed the distribution of a few diatom and dinoflagellate species over the period 1974–1983 over the same stretch of the coastal zone. At that time the sampling grid was even more refined than at present. All species listed are still present in the plankton of this area except maybe *Prorocentrum triestinum*, which showed a particular dominance during September 1975.

An increase of algal biomass in the eutrophic Dutch Wadden Sea has been described by Cadée in a series of papers (Cadée, 1990b, c, d; Cadée, 1992a; Cadée and Hegeman, 1991; Cadée, 1992a,b). One of the conclusions of his work is that there has been an increase in the period 1973–1990 of *Phaeocystis* dominance, but no such increase in diatom blooms or abundance.

When the species present in the actual situation (only diatom taxa were considered) are compared with a study from the beginning of this century (van Breemen, 1905) it is found that more than 60% of the species observed by van Breemen are still present (the difference mainly due to the genus *Chaetoceros*). Whether new species have emerged is difficult to assess due to taxonomic problems (cf. Peperzak *et al.*, in prep.).

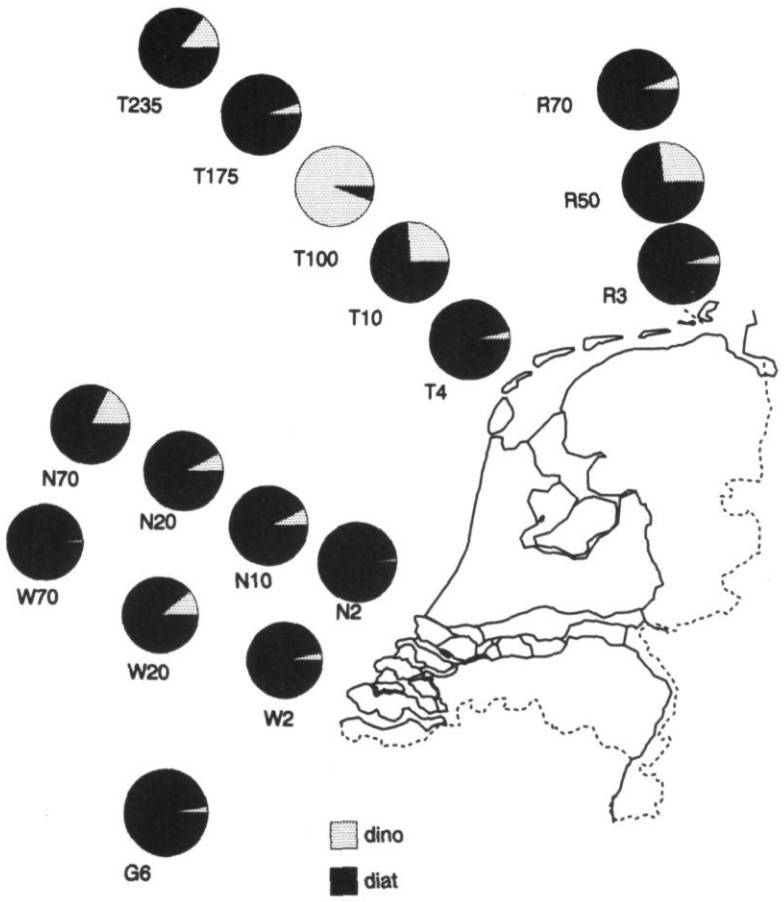

Figure 5. Mean annual ratio of cell numbers of diatoms and dinoflagellates on all stations (see Fig. 2) in 1990.

3.2. Trends in the English Channel

In a review on the changes in the phytoplankton of the Western English Channel, Boalch (1987) shows that sudden changes in primary production occurred within the long time series from 1964 until 1986. This study was based on microscopic inspection of tow-net samples taken around Plymouth. These changes occurred in the winter of 1965–1966 and were accompanied by changes in the plankton composition. In 1983–1984 a second change was observed which was accompanied by a dominance of large centric diatoms in the winter plankton. In addition to these sudden switches in the system, as shown by primary production and abundance of zooplankton or pilchard eggs, the species composition of the phytoplankton has also changed (Fig. 6). Examples of such changes are the sudden immigration or introduction of *Gyrodinium aureolum*, new species like *Halosphaera* and *Pterosperma*, the dominance during a shorter or longer time of species like *Pleurosigma planctonicum, Thalassiosira angstii* (= *T. punctigera*), and *Coscinodiscus wailesii*. The intrusion of different watermasses, depending on large-scale current phenomena, could well be one of the mechanisms behind the distribution patterns of the species, of which some show an almost reciprocal distribution, e.g. *Asterionella glacialis* and *Bacteriastrum furcatum*. Species representative of cold water obviously did well in the period 1965–1983, a cold water period. *Ceratium candelabrum* is a species indicating watermasses from the Bay of Biscay. The warm-water species *Rhizosolenia alata* fo. *indica* or *R. indica* also occurs along the Dutch coast in warm summers. Some newcomers have also disappeared again, indicating widely varying conditions for phytoplankton species over longer periods.

A further analysis of these data has been performed by Maddock *et al.* (1989). Based on a correspondence analysis they observed a high diversity throughout the year and a strong seasonal cycle. Groups of species showed a different seasonal and year-to-year occurrence. This analysis also revealed two periods (1968–1970 and 1983–1985) when changes were most obvious. Whereas these changes can be related to changes in weather patterns (cyclonic, anticyclonic, easterly, and westerly weather types), they are considered as a natural phenomenon. The changes resembled changes observed

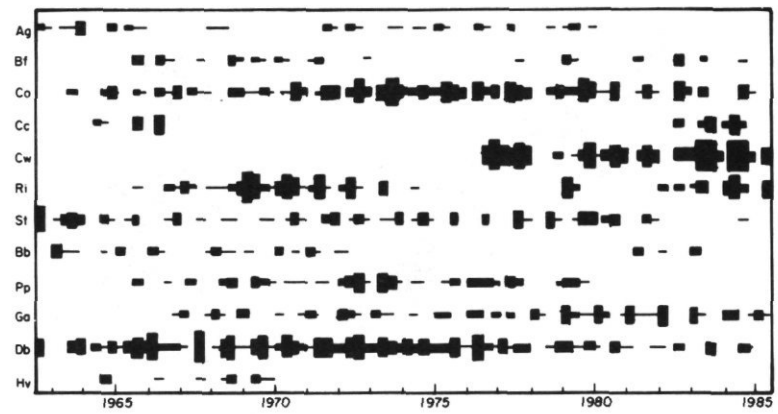

Figure 6. Occurrence of selected species of phytoplankton taken off Plymouth (Boalch, 1987) Ag, *Asterionella glacialis*; Bf, *Bacteriastrum furcatum*; Co, *Corethron criophyllum*; Cc, *Ceratium candelabrum*; Cw, *Coscinodiscus wailesii*; Ri, *Rhizosolenia alata* fo. *indica*; St, *Streptotheca thamensis*; Bb, *Braarudospharea bigelowii*; Pp, *Pleurosigma planctonicum*; Ga, *Gyrodinium aureolum*; Db, *Ditylum brightwellii*; Hv, *Halosphaera viridis*.

by other workers in other groups of organisms (zooplankton, fishes) and in other areas.

3.3 Trends in the German Bight and at Helgoland

Owing to the presence of the Biologische Anstalt Helgoland, a long time series consisting of all types of measurements (chemical, biological, physical) is available. During the last decade this series has been analysed several times with improved statistical techniques (Radach *et al.*, 1986; Radach *et al.*, 1990a,b; Radach and Bohle-Carbonell, 1990; Hickel, pers. comm.).

Hagmeier (1978) was among the first who observed an increase in phytoplankton biomass at Helgoland, which he directly attributed to an increase in nutrients. The relation between phosphate concentration and salinity suggested that this phosphate was riverine and, therefore, man-induced. The phytoplankton biomass calculated as carbon from cell counts over the period 1962 to 1990 (Fig. 7) showed an earlier increase but a stabilization in the period 1980–1990 (Hickel, pers. comm.). This latter period also shows a similar stabilization of the winter nutrient values.

The statistical analysis of the Helgoland and German Bight data, the latter taken from regular surveys in a large grid, has shown that Helgoland data are representative of a much larger area. Moreover, the data show that there is a strong cyclical mean annual course. The analysis also shows, however, that changes occur within this cyclic course: e.g. a shortening of the period with nutrient limitation, an increase of flagellates (from 2.5 μg C/l to 28.1 μg C/l) and a decrease of diatoms (from 6.2 μg C/l to 4.8 μg C/l) (Radach *et al.*, 1990b). Thus the ratio of flagellates over diatoms has strongly increased. An example of the increase in phytoplankton biomass and its seasonal cycle course is given in Fig. 8 for the period 1962–1984. Nutrient ratios have also changed owing to the increase in nitrogen and phosphate and the decrease in silicate. This decline in the Si/P ratio is held responsible for the stimulation of non-diatom blooms between 1962 and 1984 (Radach *et al.*, 1990b; cf. Smayda, 1990).

A peculiar study dealing with trends in the German Bight is one by Uhlig and Sahling (1990) on the dinoflagellate *Noctiluca scintillans*. Since 1968, Helgoland has been used as a monitoring site where seasonal periodicity with a maximum abundance in June/July and a much lower but pronounced peak in December/January has been observed. The data strongly suggest a three-year cycle in abundance of *Noctiluca*. Since 1981 the German Bight proper is included in the studies; these show that mass aggregations are observed at the sea surface which are often misinterpreted as blooms. In fact, they are the final moribund stadium of previous blooms of this organism, which generally vanish within a few days with intensive wind-induced mixing. Although *Noctiluca* is often treated as a phytoplankter, one should realize that it is a heterotrophic dinoflagellate and therefore could well react completely differently from other autotrophic phytoplankters.

A study by Bätje and Michaelis (1986) on *Phaeocystis* in the German Wadden Sea has suggested that *Noctiluca* closely follows the *Phaeocystis* blooms. As stated before, Owens *et al.* (1989) have also observed a three-year cycle in the occurrence of *Phaeocystis* in the Northeast Atlantic. It would be interesting to investigate whether a causal relationship exists between these cycles.

3.4. Trends in the Baltic Sea

Based on a review by Gerlach (1988), a recent paper (Anon. 1990a) of the Baltic assessment group, a special volume of Ambio (Anon. 1990b) and a paper by Niemi

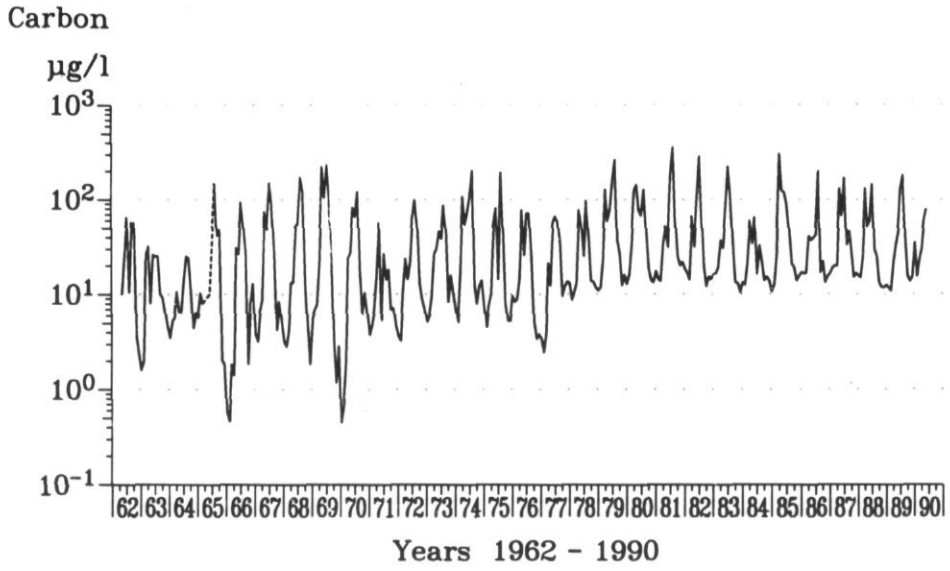

Figure 7. Monthly median biomass of phytoplankton at Helgoland (Hickel, unpubl., pers. comm.).

(1982), we have a wealth of information on the situation in the Baltic proper and in several subareas. Winter values of nutrients obviously have increased several times, with phosphate up to two to three times higher, and nitrogen three to four times. Also, biomass of phytoplankton in summer has increased when 1975–1978 and 1980–1983 are compared (Gerlach, 1988; Nehring, 1989; Wulff and Rahm, 1988).

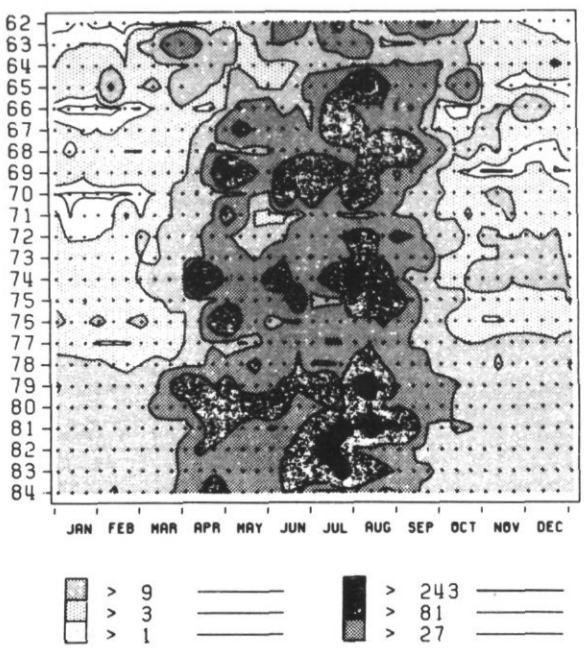

Figure 8. Annual cycles of phytoplankton at Helgoland Reede, presented as 10-day means (Radach *et al.*, 1990a).

In the subareas, such as the Archipelago Sea (Viitisalo *et al.*, 1990) chlorophyll has increased concomitantly with higher nutrient levels in winter. Salinity has decreased by 0.5. Some zooplankton species followed the decrease in salinity. Owing to the high interannual variability, however, eutrophication effects in the open sea are generally still moderate. It proves to be difficult to discriminate between direct anthropogenic inputs (rivers, atmosphere) and indirect nutrient inputs from deep water. The period 1984–1988 shows an increasing tendency for chlorophyll in most subareas of the Baltic Sea. Primary production shows a parallel tendency. Despite the observation of exceptional and, in some cases, toxic blooms no tendency for large-scale changes in phytoplankton species composition has been found.

Semi-enclosed areas and bays with a limited circulation and mixing obviously may behave differently: a study by Grönlund and Leppänen (1990) in the western Gulf of Finland has shown that peaks of spring phytoplankton blooms are more intensified as a result of the increase of nutrient reserves during the period 1972–1988 (Fig. 9).

Niemi (1982) has described plankton blooms in the Central and Northern Baltic. An assessment of the present situation has been presented in Anon. (1990a) for the whole Baltic Sea area. Their conclusions are comparable to those of Gerlach and the reader is referred to the summary in Anon. (1990a).

A study of the annual cycle of Kiel Bight plankton was made by Smetacek (1985), who concludes that variations in species composition are not linked to long-term trends over the past century, in spite of the large increase in anthropogenic nutrient inputs. Short-term events of a physical (mixing and stratification) and a biological

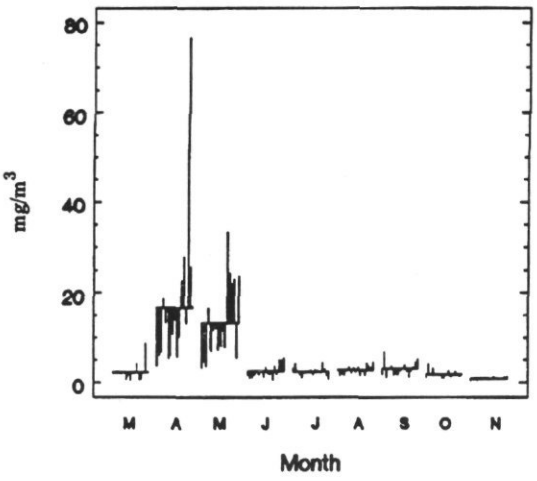

Figure 9. Concentrations of chlorophyll *a* per month. Horizontal lines represent monthly averages during the period March–November in 1972–1988. Each vertical line represents one year and is plotted from the long-term average to the monthly mean concentration of the respective year (Grönlund and Leppänen, 1990).



Figure 10. Phytoplankton blooms based on Mommaerts (1985).

(grazing) nature are far more important in determining the occurrence of fast-growing species.

Of great importance for predicting future developments of the Baltic Sea is knowledge of the mechanisms which promote upwelling and mixing with deep nutrient rich water (Anon., 1990a). The very limited exchange of the Baltic Sea with the North Sea causes the Baltic Sea to behave quite independently of the North Sea and North Atlantic.

3.5. Phytoplankton blooms

To gain insight into the occurrence of blooms in the ICES area all the data compiled by Mommaerts (1985, 1986a, b) have been analysed and partly mapped, showing phytoplankton blooms until 1982 (Fig. 10). He also lists species with a recurring character (Mommaerts, 1986a) and all blooms occurring in 1984 are listed in Mommaerts (1986b). New data taken from several sources are given in Figure 11 and listed in Table 1, covering the period until 1990. A striking result of the representation of blooms in these maps is that there is a close relation between areas where blooms are observed and the vicinity of scientific institutes: main areas of interest are the Scandinavian coastal waters with institutes in Oslo and Bergen; the Skagerrak and Kattegat with institutes in Denmark and Sweden (Lund and Kristineberg); the Baltic Sea with institutes at Tvarminne and Helsinki; the German Bight and the Dutch coastal zone with institutes at Helgoland, Hamburg, and Texel; the Channel with institutes in Plymouth (UK) and Roscoff and Brest (F). The Irish Sea and the southern Irish coast are covered by institutes in Bangor and on Sharkin Island and the northern part of the United Kingdom by institutes in Scotland (Aberdeen and Oban). Finally, Vigo and Porto are main centres in the south. Thus observer effort, mentioned as one of the possible cases for the increase of algal blooms (Smayda, 1990) is really one of the factors to be dealt with. Also it is observed that sampling is biassed towards coastal regions which are much more accessible. This is surely a point in favour of the CPR studies which, however, seldom sample appropriately the coastal zones (Radach *et al.*, 1990a; Cadée, 1990a).

Other conclusions can be extracted from these figures: there are several areas where blooms show a recurring pattern (Figs. 10 and 11). This pattern is probably both natural as well as influenced by eutrophication. It would be very useful to continue the work started by Mommaerts within ICES to be able to answer questions such as: has the frequency of (toxic, nuisance) blooms increased, or, which species are the most important in bloom formation? This could also shed light on the question whether these blooms are of natural or of anthropogenic origin.

The map presented by Partensky and Sournia (1986) has been supplemented with data of *Gyrodinium aureolum* from the Dutch coast, where it has been traced in most marine and brackish waters (Klein and van Buuren, 1992; Koeman *et al.*, 1991; Zevenboom *et al.*, 1990). It must therefore be concluded that Smayda's confidence in Dutch planktonologists was not justified. This leaves his remark on the effects of toxic compounds in river waters unimpeded (Smayda, 1989, p. 455).

Table 1. Phytoplankton blooms (compilation based on references listed), as shown in Figure 11.

No.	Species	Area	Author
1	*Gyrodinium aureolum*	Skagerrak	Lindahl, 1986
2	*Chrysochromulina polylepis*	Kattegat	Nielsen *et al.*, 1990
3	*Ceratium furca*	Helgoland	Hickel, 1982
4	*Phaeocystis* spp.	North Sea	Reid *et al.*, 1990
5	*Corymbellus aureus*	North Sea	Gieskes and Kraay, 1986
6	n.g., Chlorophyll	Baltic Sea	Kahru *et al.*, 1990
7	Dinoflagellates	SW Sweden	Granéli *et al.*, 1989
8	*Phaeocystis* spp.	Wadden Sea	Cadée, 1990d
9	*Noctiluca* sp.	Wadden Sea	Cadée, 1990b
10	*Chrysochromulina birgeri*	SW Finland	Niemi, 1982
11	Diatoms, *Gonyaulax*	SW Finland	Niemi, 1982
12	*Cyanobacteria*	SW Finland	Niemi, 1982
13	Marine centric diatoms	SW Finland	Niemi, 1982
14	*Heterocapsa triquetra*	SW Finland	Niemi, 1982
15	*Prorocentrum minimum*	Oslo fjord	Smayda, 1990
16	*Rhizosolenia delicatula*	Dutch coast	Reid *et al.*, 1990
17	*Skeletonema costatum*	Dutch coast	Reid *et al.*, 1990
18	*Gyrodinium aureolum*	Channel	Reid *et al.*, 1990
19	*Distephanus speculum*	Kattegat	Reid *et al.*, 1990
20	Coccolithophores	North Sea	Holligan, 1987
21	*Mesodinium rubrum*	Dutch coast	Koeman *et al.*, 1991
22	*G. aureolum, G. tamarensis*	W. Scotland	Gowen, 1987
23	*Phaeocystis* sp.	Channel	Boalch, 1987
24	*Noctiluca scintillans*	Channel	Boalch, 1987
25	*Phaeocystis* sp.	Barents Sea	Wassmann *et al.*, 1990
26	*Noctiluca scintillans*	North Sea	Uhlig and Sahling, 1990
27	*R. delicatula, L. danicus*	Dutch coast	Kat, 1988
27	*P. balticum, P. minimum*	Dutch coast	Kat, 1988
28	*Gyrodinium aureolum*	Dutch coast	Zevenboom *et al.*, 1990
29	*Phaeocystis* sp.	Irish Sea	Rogers and Lockwood, 1990
30	*Phaeocystis* sp.	North Sea	Lancelot *et al.*, 1990
31	*Noctiluca scintillans*	North Sea	Lancelot *et al.*, 1990
32	*Ceratium* spp.	N. North Sea	Dickson *et al.*, 1992
33	Cf. *Lepidodinium viride*	German Bight	Elbrächter *et al.*, in prep.
34	*Prorocentrum minimum*	Baltic Sea	Jochem and Babenerd, 1989

Based on the CPR the spatial and temporal distribution of blooms has been analysed in the Northeastern Atlantic and the North Sea (Reid *et al.*, 1987). These analyses show that blooms of diatoms and *Ceratium* spp. have declined over the last 26 years (1958 to 1983), but also that there is some evidence of an increase since 1975 of high levels of phytoplankton, based on the green colour of the recorder silks.

These authors also touch on the problem of "what is an exceptional bloom?" Some of the criteria used to distinguish exceptional blooms are: they occur in previously unrecorded areas; the species forming the bloom are new to the area; the timing is different from normal with respect to area or species; they are more extensive than usual; they have some deleterious effect, such as PSP or other toxic phenomena. Of all samples analysed in the North Sea between 1958 and 1983, about 4 to 5% showed "normal" blooms. The species causing these blooms have been listed in Reid *et al.* (1987, his Table 1) with *Thalassiosira* spp. and *Ceratium fusus* as the most frequent bloom-forming genera for diatoms and dinoflagellates, respectively. Others, however (cf. Tett, 1987; Cadée, 1990d; Holligan, 1987), used other criteria to describe blooms, such as cell densities ($>10^6$ cells/l) and chlorophyll *a* concentration. Holligan (1987), although aware of the subjective character to define criteria of exceptional blooms, argues that spatial and temporal dimensions of >10 km and of >10 days are probably very important. Those blooms where chlorophyll exceeds >300 mg m^{-2} must be considered exceptional.

A particular species studied intensively during the last five years is *Phaeocystis*. The scientific and political interest was stimulated by the occurrence of large foam masses on the beaches of the countries bordering the Southern Bight of the North Sea, such as The Netherlands, Germany, and Denmark. A programme including many aspects of this species, including distribution, physiology, and ecology, is subsidized by the EC with many countries around the North Sea cooperating. Recent information has been published in Lancelot *et al.* (1990). Data for the Marsdiep by Cadée have shown that since 1970 a steady increase in the growth period of *Phaeocystis* has occurred, up to 130 days per year. Older data from the end of the last century show that at that time *Phaeocystis* was a dominant phytoplankter for up to

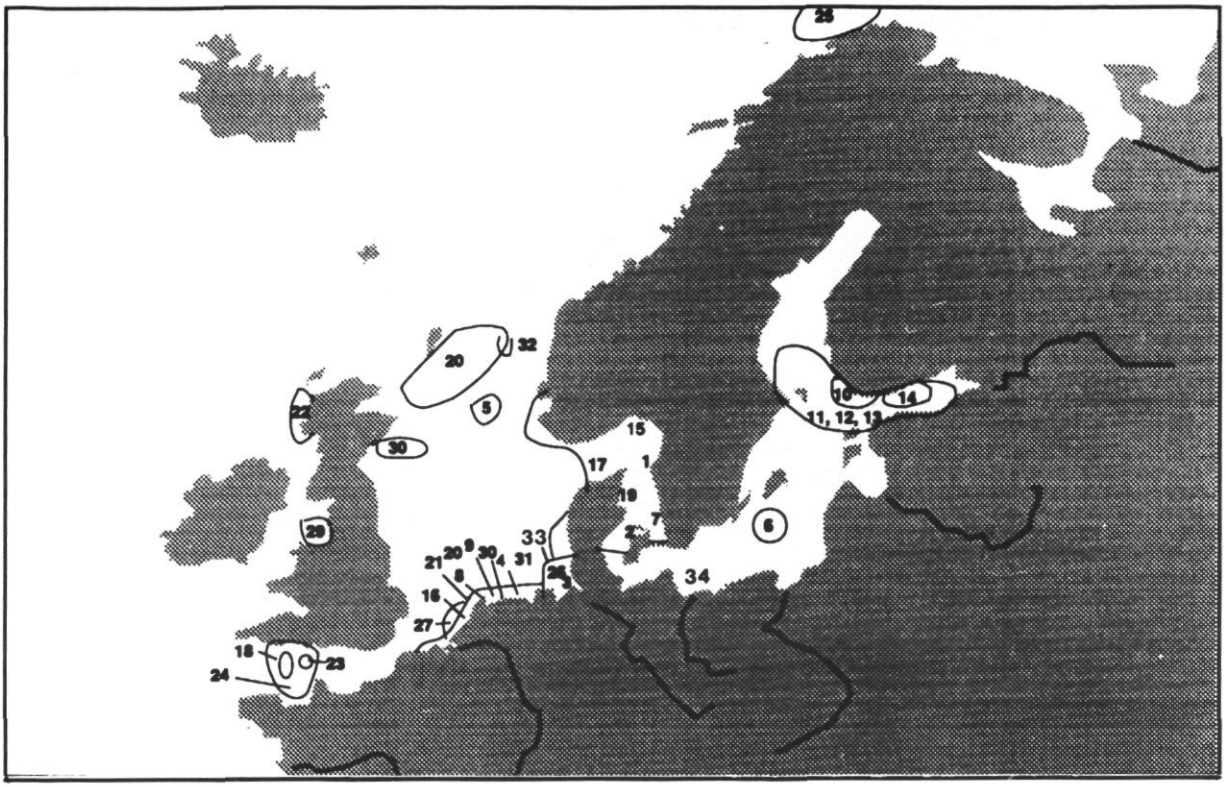

Figure 11. Areas with phytoplankton blooms compiled from several papers; numbers refer to Table 1.

40 days per year (Cadée, 1990d). The increase of *Phaeocystis* blooms has been ascribed almost exclusively to the nutrient concentrations in the coastal watermasses. As shown in Figure 11, however, blooms of *Phaeocystis* are normal components of the seasonal succession pattern in other areas, such as the Barents Sea (Wassmann *et al.*, 1990) In the Antarctic region as well, blooms of *Phaeocystis* are ubiquitous (Smetacek *et al.*, 1991). It is uncertain whether all these observations concern the same species (cf. Buma *et al.*, 1991). The dominance of species such as *Phaeocystis* can only be understood after careful examination of its growth and adaptive capacities in terms of resource competition. Studies by Riegman *et al.* (1992) show that *Phaeocystis* competes well with other species at specific nutrient ratios, especially nitrate and ammonium.

A bloom-forming species which attracted very much attention in 1988 was *Chrysochromulina polylepis* (Lindahl and Dahl, 1990; Skjoldal and Dundas, 1991). Owing to its toxic behaviour, several studies were performed to find out why the bloom occurred and why the species became toxic. Scientists competed to give fanciful explanations for the sudden appearance and the toxic character of the bloom, such as high winter nutrient stocks, growth stimulation from selenium, growth stimulation from cobalt, specific weather conditions, and possible disturbances of the plankton community resulting from herbicides or even toxic substances originating from the Rhine (Granéli *et al.*, 1989).

In conclusion, despite the huge amount of information, no clear evidence for an increase in bloom events in the ICES area in general can be given. Locally, areas can be found where changes have occurred. Some of these changes were temporary, others seem fairly permanent.

3.6. Algal succession

Seasonal succession patterns have been analysed in search of changes in marine and estuarine systems: a few studies will be touched on. The effects of the construction of the storm surge barrier in the Dutch Delta region have been followed in detail. In one of the many studies of this ecosystem (Bakker *et al.*, 1990), seasonal succession of phytoplankton in the Eastern Scheldt has been evaluated, before and during the construction, using a canonical correspondence analysis. During the period investigated, the transparency has increased due to the reduced current velocities. This improvement in the underwater light climate induced a change in the phytoplankton assemblage towards a summer character, extending its growth season both earlier and later in the year. The light factor explained not only the seasonal

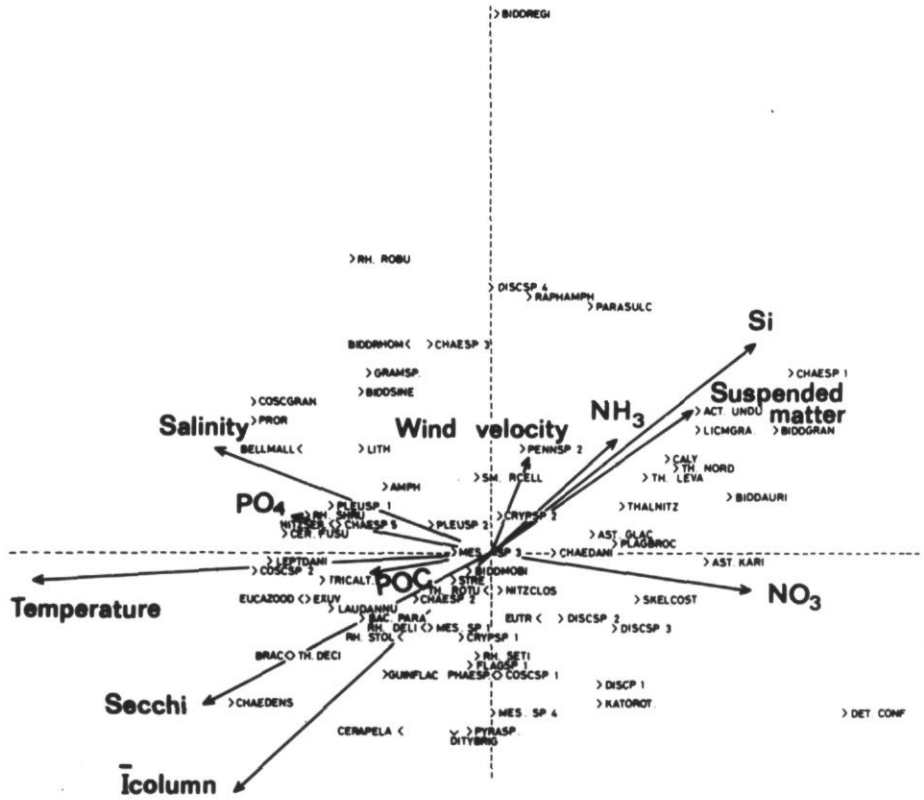

Figure 12. Canonical correspondence analysis of the phytoplankton in the eastern part of the Oosterschelde: scores of environmental variables and species in the plane of the CCA ordination. For species names and explanation, see Bakker *et al.* (1990).

pattern, but also the long-term trend from pre-barrier to barrier period (Fig. 12).

Phytoplankton succession in the Channel (Maddock *et al.*, 1989) was studied using correspondence analysis. These analyses show that a strong seasonal cycle with systematic changes in species composition occurs. Depending on the quantitative presence of groups of species the seasonal cycles show shifts over longer periods which could be related to weather conditions, e.g. the effect of enhanced mixing in winter causes benthic diatoms to predominate.

Succession patterns in phytoplankton along the Dutch coast have been reported by Gieskes and Kraay (1975), Leewis (1985), and Kat (1982). In this volume, Cadée reports on the succession of a series of selected phytoplankton species in the Marsdiep. Unfortunately the Helgoland data have not yet been treated at the species level, although the data are available. The succession pattern of the phytoplankton in the Kiel Bight is described by Smetacek (1985) in general terms with an indication of the dominant species.

4. Causes of blooms

Because of their dependence on light and nutrients, the causes of phytoplankton blooms have mostly been related to two complexes of factors which cannot be discriminated completely owing to their intricate coupling. In this paper physical factors will be treated first because of their ubiquitous and overwhelming influence on phytoplankton growth. Physical factors include (global) irradiance, turbidity, turbulence, and tidal mixing. Eutrophication includes the effects of anthropogenic inputs of nutrients. These effects are concentrated in those areas where inputs are high through riverine transport. Atmospheric deposition will not be dealt with, although this factor might become more important in the future and in remote areas not under the direct influence of river water. The possible role of micronutrients, such as trace metals and organic compounds, will not be discussed. As a third cause, biological and other interactions will be touched on. Reduced grazing as a consequence of contaminants could be an explanation for observed eutrophication effects.

4.1. Physical factors

Holligan (1987) has described the basic controlling factors for phytoplankton growth and succession: turbulence versus stratification or stabilization of watermasses

Table 2. Normal phytoplankton succession with time (Holligan, 1987).

Physical state of environment	Turbulent ⟶ Stable
Dominant phytoplankton	Diatoms ⟶ Flagellates ⟶ Dinoflagellates
	r-selected ⟶ K-selected (Pianka, 1970)
Life strategy	(Productive species with short division times) ⟶ (Efficient species with long division times)
	Ruderals ⟶ Competitors ⟶ Stress-tolerators (Grime, 1979)

and the availability of light and nutrients (Table 2). Different growth and reproductive strategies, present in different groups of phytoplankton (Margalef, 1978), are linked with characteristics such as growth rate, cell size, the ability to form colonies, nutrient uptake rates and nutrient affinities, photoadaptation, and sinking rate. Interaction of these biological characteristics with physical and chemical factors is difficult to resolve. At the onset of phytoplankton blooms, at least at temperate latitudes, irradiance and mixing depth are most important. Subsequently, for the seasonal succession the physical factor of overriding importance is the temporary transition of a turbulent mixed water column to a stratified stable column.

Situations which may lead to a physical environment of exceptional blooms are described by Holligan (1987) for diatoms, flagellates, and larger dinoflagellates. They all are induced by more or less specific conditions. For example, diatoms develop exceptional blooms with low turbidity conditions, with abnormal salinity stratification, nutrient pulses both from rivers and by advective processes, and with wind-induced summer upwelling. Examples of favourable conditions for other groups are illustrated as well (cf. Holligan, 1987).

Several other authors have given examples of the influence of physical factors on plankton dynamics. Lindahl (1986) and Lindahl and Hernroth (1988) showed that changes in phytoplankton and zooplankton densities were caused by inflow of high salinity surface water from the Skagerrak into Gullmarsfjorden. This water contained e.g. *Gyrodinium aureolum* growing offshore (Fig. 13). Within the fjord concentration pro-

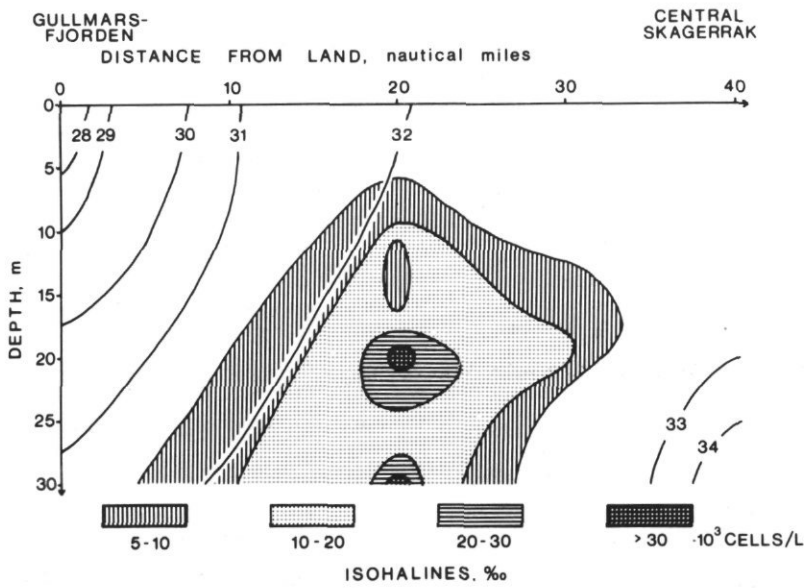

Figure 13. Vertical and horizontal distribution of *Gyrodinium aureolum* and salinity along a transect westwards from Gullmarsfjorden on 28 and 29 August 1983 (Lindahl, 1986).

cesses occur causing high densities of this species but also of zooplankton.

In their analysis of the Helgoland data Radach and Bohle-Carbonell (1990) showed that most meteorological and oceanographic parameters exhibited unchanged annual cycles within natural variability, except for sea-surface temperature, which has increased about 1°C in the last 23 years. There was no indication of a possible effect of this temperature rise on phytoplankton composition, probably due to the dominating influence of chemical factors in this area (see next section).

Cadée (1992a) concluded that a high winter temperature did not influence the succeeding timing of the phytoplankton spring bloom: more light and clear water, however, caused peaks to occur earlier.

The spring bloom in the Baltic Sea has been studied in detail by Kahru and Nommann (1990), who showed that this bloom is triggered by saline stratification and light rather than by the establishment of thermal stratification. Moreover, the bloom progresses from the more stratified areas towards the less stratified central areas.

Causes of the exceptionally early blooms of *Ceratium* spp. in the southern North Sea have been observed to coincide with changes in meridional circulation with strengthened southerlies during the early 1970s (Dickson and Reid, 1983). In conclusion, the importance of physical factors in regulating (exceptional) blooms is clear, but depends on the sea area and local conditions (depth, vertical salinity distribution, and mixing). The distinction between eutrophication and physical factors, however, is often hindered (see Gillbricht, 1983 and next section).

4.2. Eutrophication

The discussion on algal blooms in the North Sea and the Baltic and their connections (Skagerrak, Kattegat) is strongly associated with eutrophication (Gerlach, 1988; Anon. 1990a; Grönlund and Leppänen, 1990; Andersen, 1986; Smetacek, *et al.*, 1991; Tämminen and Leskinen, 1985; Granéli *et al.*, 1989; Granéli and Granéli, 1989; Granéli *et al.*, 1990; Gieskes and Schaub, 1990; Schaub and Gieskes, 1991; Kat, 1982; Kat, 1988; Lancelot *et al.*, 1990; Cadée, 1990a, c, d; Peeters *et al.*, 1990; Peeters and Peperzak, 1990; Radach *et al.*, 1990a, b; Klein and van Buuren, 1992). From these papers one can conclude that increasing nutrient levels induce enhanced algal biomass. Examples of such cases are observed along the Dutch coast, where watermasses appear with different nutrient concentrations and a concomitant gradient in algal biomass (Schaub and Gieskes, 1991; Klein and van Buuren, 1992; Fransz and Verhagen, 1985). Studies in the German Bight have also shown that the increase in nutrient levels has not only stimulated algal growth (Bauerfeind *et al.*, 1990), but also a change in algal species composition (Radach *et al.*, 1990a, b). In enclosed water bodies, such as bays with a limited water circulation and high riverine inputs, similar observations have been made (Laholm Bay, Sweden and specific subareas in the Baltic Sea) (Edler, 1984; Anon., 1990a). Bioassay experiments have shown the predominant influence of nutrients in the North Sea (Peeters *et al.*, 1990) and the Baltic Sea area (Granéli *et al.*, 1990).

For biological effects of eutrophication the reader is referred to Hickel *et al.* (1989), Rachor (1985), von Westernhagen *et al.* (1986), and Rogers and Lockwood (1990).

Apart from this generalized view, the occurrence of species-specific blooms is much more difficult to attribute directly to nutrient levels *per se*: there is growing evidence that not only nutrient concentrations but much more the ratios in which these nutrients are advected are a cause of the successful competition of these species (e.g. *Phaeocystis*) (Riegman, 1991). In a recent paper by Tett and Mills (1991) the circumstances in the mixed and turbid southern part of the North Sea are set against the growth conditions in the Northern part (stratified, clear water). They assume that the Northern part is a stable system with zooplankton adapted and temporarily fitted to the availability of food derived from the vernal blooms, whereas in the southern part the system can be regarded as less stable as far as the tight coupling between phyto- and zooplankton is concerned (cf. Fransz and Gieskes, 1984; Cushing, 1989). However, long-term series to support their observations are few.

Nutrients and algal blooms along the Dutch coast in the period 1975–1990

One of the large data sets of older monitoring studies along the Dutch coast has been analysed recently to study the possible trends in nutrient concentrations in relation to algal blooms as chlorophyll *a* concentrations. A statistical analysis of the nutrient data has shown (Klein and van Buuren, 1992) that only slight but significant trends can be observed in Dutch coastal waters. However, these trends are both upward and downward, depending on chemical compounds and locations. The concentrations of ammonium showed decreasing trends close to the Rhine plume, whereas offshore significant upward trends were observed. Nitrate did not show any trend and was stable over the 15 year period. The same holds for silicate. Phosphate concentrations showed a downward trend close to the coast, whereas offshore a twofold increase was observed. To date, these trends are not yet understood, although the average winter concentrations of nitrate and phosphate have increased three and two times respectively in the period 1986–1988 as compared to the values measured around 1930 (Sydow *et al.*, 1990; Kremling and Pohl, 1989).

For the purpose of description of blooms we have developed another criterion for the assessment of the occurrence of blooms: differences from the annual

median value of chlorophyll *a* have been plotted. The large differences in absolute concentrations between coastal and offshore stations are better represented in this way. For a series of years these values are presented in Figure 14 for three stations along the Noordwijk transect showing a large interannual variation in both the timing and frequency of blooms over the years and along the transect.

It is self-evident that at least the density of these blooms is related to the nutrient levels in the coastal zone. The gradients in concentrations of the nutrients (silicate, phosphate, and nitrogen compounds) are very obvious, with high concentrations in winter close to the coast and low concentrations offshore, and a rapid decline in concentrations of all components during the growing season of phytoplankton (these data are presented in Klein and van Buuren, 1992).

In several studies (Peeters and Peperzak, 1990; Riegman *et al.*, 1990; Peeters *et al.*, 1990; Bauerfeind *et al.*, 1990) attempts have been made to find out which nutrients limit the phytoplankton growth in this area. In general these studies show that silicate limits diatom growth both in the nearby coastal zone and offshore, whereas phosphate limits growth along the coast and nitrogen offshore, apart from the limiting role of irradiance in winter. However, evidence is growing that also nutrient ratios and in particular nitrate/ammonium ratios influence the species composition of phytoplankton (Riegman, 1991).

In his excellent review on the role of competition for resources in phytoplankton succession, Sommer (1989) gives clues to the importance of nutrient ratios and nutrient/light ratios for algal succession. Evidence from culture experiments, mainly with freshwater algae, has

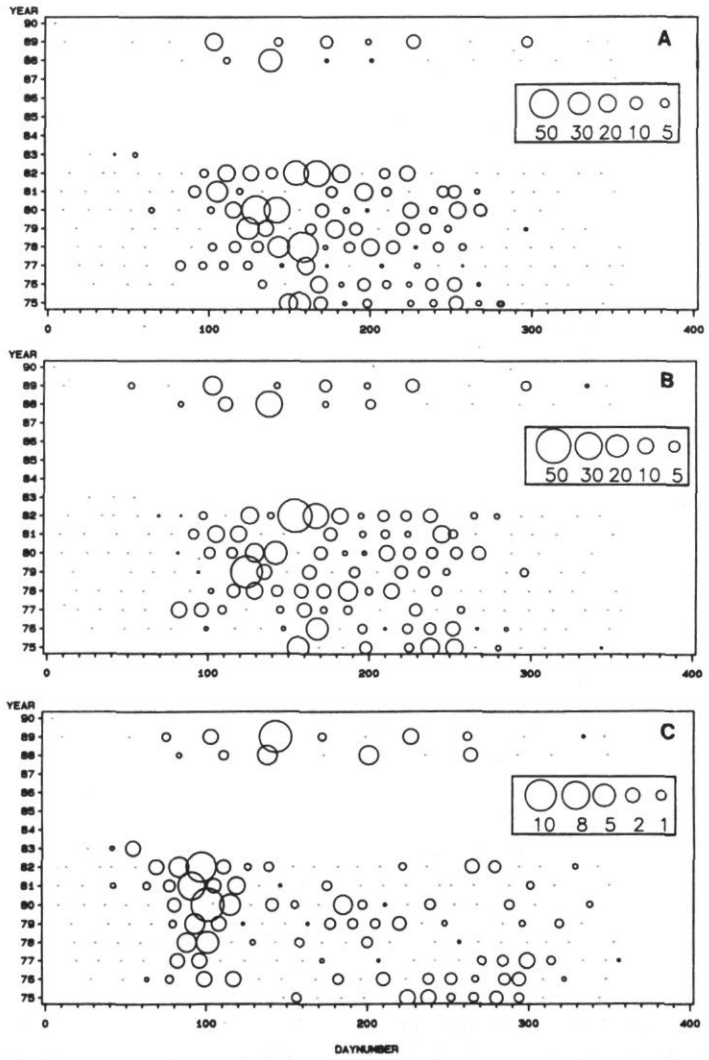

Figure 14. Chlorophyll *a* values minus median values at station N-2 (a), at station N-20 (b), and at station N-70 (c). Median for N-2 is 4.3, for N-20 2.8 and for N-70 1.4 μg chlor.a/l, respectively.

shown that high Si/P ratios induce dominance of diatoms, and that low N/P ratios induce predominance of cyanobacteria. There is a great need for more ecophysiological information regarding marine phytoplankton species.

4.3. Biological and other factors

Leymann (1988) has suggested two alternative causes for the expression of eutrophication effects: an observed increase in tidal mixing due to the enhanced tidal amplitude in the German Bight and the increase of the input of EDTA like compounds. Experiments with the latter suggest that phytoplankton growth increases after the addition of EDTA, a complexing agent. Granéli and Moreira (1990) showed that humic rich river waters mixed with sea water stimulates the growth of dinoflagellates in mesocosm experiments and also provokes a shift from diatoms to dinoflagellates.

In a study on the relations between aluminium and phytoplankton growth van Beusekom (1991) argues that a close link exists between solubility of diatom frustules and their aluminium content. Therefore he presented (1991) an alternative hypothesis, as opposed to suggestions made by Colebrook *et al.* (1984) and Dickson *et al.* (1988) to the decrease of diatom blooms in the North Sea. His experiments with diatom cultures showed that frustules grown under enhanced aluminium concentrations show a retarded silicate remineralization compared with frustules grown under low aluminium concentrations. Assuming a higher aluminium input into the North Sea and an increased aluminium flux from the sediments due to higher mineralization caused by more organic matter, the decrease of diatoms (Radach *et al.*, 1990a, b) could be explained.

An example of the effect of reduced grazing and subsequent development of the *Chrysochromulina* bloom is given by Kiorboe *et al.* (1990), Nielsen *et al.* (1990), and Carlsson *et al.* (1990). Carlsson *et al.*, suggests that the elimination of grazers was one of the mechanisms causing the intensity of the *Chrysochromulina* bloom. Also natural blooms may be controlled by grazing as shown by Admiraal and Venekamp (1986) (Fig. 15), Bathmann *et al.* (1990), Weisse and Scheffel-Möser (1990), whereas others (Wassmann *et al.*, 1990) have shown that blooms may sediment out of the photic layer unused by grazers.

In conclusion, we are still far from understanding the wax and wane of blooms. Modelling of the principal processes has just started (Thingstadt and Sakshaug, 1990) and will only progress when results of field measurements and experimental data of cultures and experiments in the field and mesocosms are combined. Owing to the importance of physical processes, three-dimensional modelling of water transport will play a key role in the prediction of phytoplankton growth and blooms.

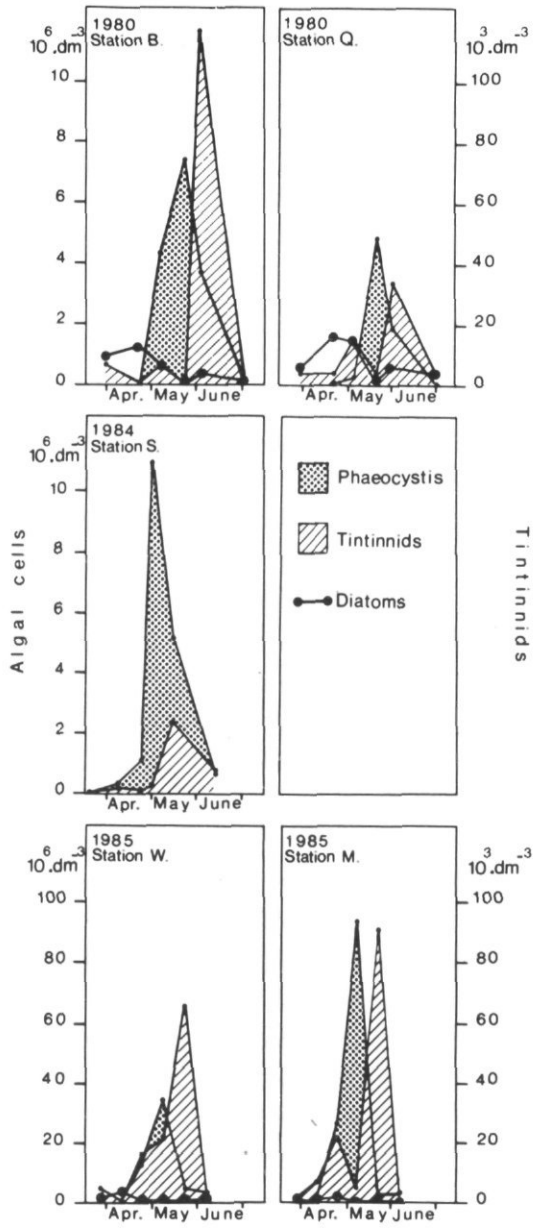

Figure 15. Effect of tintinnid grazing on biomass and density of *Phaeocystis* cell numbers (Admiraal and Venekamp, 1986).

5. Conclusions and recommendations

Despite the amount of information on algal blooms, generally there is no clear evidence of an increase in bloom events in the ICES area. Locally, areas can be found where changes have occurred, some of them temporary, others fairly permanent.

Exceptional phytoplankton blooms can be observed almost everywhere in the ICES area. Observation of

blooms is strongly influenced by observer effort. Causal factors for the occurrence of blooms are of a physical and chemical nature. Eutrophication may override the effects of physical factors in areas of limited water circulation or mixing. Public awareness forces scientists and policymakers to investigate blooms and to explain their occurrence (cf. Smetacek, 1991).

The following recommendations can be made:

(1) There is still an urgent need for standard methods for studying phytoplankton (cf. Niemi *et al.*, 1985): (a) to improve methods for counting and identification, (b) to study phytoplankton at relevant temporal and spatial scales.
(2) Criteria have to be developed which can be applied generally to describe blooms, exceptional blooms, growth, or specific densities of algae.
(3) CPR data are useful but data on the decrease of diatom species do not fit into the observed increase in phytoplankton colour, which is tentatively attributed to flagellates. If possible, the CPR should be improved to study adequately small phytoplankton species.
(4) There is a need for a databank on phytoplankton species composition at a few selected sites, intercomparison of results is needed using working groups like those active under COST and ICES. The stations used up to now are most suited to extend the already available series: Plymouth, Texel, Helgoland, Kiel, SW Finland. Station N-10 of the Dutch monitoring network is an appropriate location for the Dutch coastal zone. To reach this goal governmental agencies should guarantee long-term financial resources.
(5) Lack of autecological knowledge of phytoplankton growth in relation to nutrient ratios makes a prediction of blooms still almost an impossibility. An understanding of bloom-inducing mechanisms should be studied in much more detail. A combination of such studies with long-term monitoring is needed to be able to give up-to-date information both to the public and to the policymakers (ten Brink *et al.*, 1991).

Acknowledgments

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Phytoplankton variability in the Marsdiep, The Netherlands

Gerhard C. Cadée

Cadée, G. C. 1992. Phytoplankton variability in the Marsdiep, The Netherlands. – ICES mar. Sci. Symp., 195: 213–222.

Phytoplankton has been sampled regularly in the Marsdiep, the westernmost tidal inlet of the Wadden Sea, since the early 1970s. During this period, *Phaeocystis* was numerically the most important species. Duration of blooms of non-diatoms, particularly of *Phaeocystis* (defined as periods with >1000 cells · ml^{-1}) have increased since the early 1970s from ~30 to >100 days. In the 1980s, duration of *Phaeocystis* blooms remained ~100 days, with highest values at the end of the decade (120–150 days), a continuing increase not observed in other non-diatom plankton. The duration of diatom blooms fluctuated considerably, but at the end of the 1980s they were the same as at the end of the 1960s. Average annual phytoplankton biomass (chlorophyll *a*) showed a good correlation with annual average Rhine runoff measured at Lobith. This indicates the importance of nutrients carried by Rhine water either reaching the Marsdiep via the River IJssel or Lake IJssel or with the residual coastal current along the Dutch coast. Enhanced leaching of nutrients particularly N from over-fertilized agricultural land during high rainfall seems the main cause of the observed correlation of phytoplankton biomass and algal blooms in the Dutch coastal waters.

Gerhard C. Cadée: Netherlands Institute for Sea Research, PO Box 59, 1790 AB Den Burg, Texel, The Netherlands.

Introduction

The knowledge of natural variability of the ecosystem is important when studying trends due to man-induced changes of the environment, e.g. eutrophication or global warming. Phytoplankton variability has been studied in a broad spectrum of scales ranging from tidal (e.g. Veldhuis *et al.*, 1988), diurnal (e.g. Sournia, 1975), and neap-spring tidal cycles (Balch, 1986) to annual (e.g. Cadée, 1986a), and year-to-year scales (e.g. Tont, 1987).

One may question whether a time period of ten years (1980–1989), as chosen for this symposium, is long enough to illustrate long-term variability. It almost coincides with the 11-year cycle in sunspots: maximum sunspot activity coinciding with nice summers and more severe winters in our area. The sunspot cycle is prominent in tree-ring records (Zeuner, 1958). Ottestad (1942) could correlate cod catches with tree-ring width and thus sunspot cycle in northern Norway. Vibe (1967) observed 11-year sunspot cycles in among others drift-ice extension, cod landings, and numbers of polar bears, harp seals, and ptarmigans in Greenland. Colebrook and Taylor (1984) observed plankton periodicity with a wavelength of 10–12 years which, however, they did not relate to a particular process.

Variations in plankton with cycles of less than 10 years are known: cycles of about three years are present in the data of the Continuous Plankton Recorder (CPR) survey (Colebrook and Taylor, 1984; Colebrook, 1986). El Niño with its marked effects on plankton and fisheries off Peru, but also elsewhere, occurs about every four years. The interval between two successive events varies from 2 to 10 years (Barber and Chavez, 1983; Cane, 1983).

Long-term changes in climate of about 50 to 70 years' duration express themselves in periodic changes in wind strength and direction (Cushing, 1975). Cushing stresses the influence on marine phytoplankton production cycles and hence on the match or mismatch of fish larval production and consequently fish recruitment. The CPR survey data for the North Sea (since 1931) and for the Northwest Atlantic (e.g. Colebrook, 1986; Owens *et al.*, 1989) and the so-called Russell cycle observed in plankton of the English Channel since 1924 (Southward, 1963; Russell, 1973) are the best data available for illustrating such long-term trends. The existence of such long-term cycles equalling the lifetime of a scientist makes the study of natural vs. man-induced variations in the sea no easier (Gillbricht, 1986). This implies that the debate whether changes observed in (phyto)plankton are due to climate or eutrophication will continue.

This paper summarizes data on phytoplankton variability in the Marsdiep, the westernmost tidal inlet of the Wadden Sea. Monitoring phytoplankton started in the 1970s and historical data going back to 1897–1899 (Cleve, 1898–1900; Cadée and Hegeman, 1991) can be used for comparison. Earlier reports on this data set have been presented (Cadée, 1986a, 1990b, 1992; Cadée and Hegeman, 1991). Phytoplankton changes observed in this coastal area are comparable to those found in the Rhine-influenced part of the Southern Bight, North Sea (Gieskes and Kraay, 1977), and in the Helgoland series (Radach *et al.*, 1990): increase in non-diatom phytoplankton and no increase in diatoms. They are different from changes observed in the open North Sea (e.g. Owens *et al.*, 1989), indicating that climatic changes effective in the open North Sea are overridden by man-induced changes in nutrient (ratios) in the continental coastal North Sea waters (van Bennekom *et al* 1975); Cadée, 1990a; Riegman *et al.*, 1992a,b).

Methods

Methods are treated more *in extenso* in earlier papers (Cadée, 1986a, 1992; Cadée and Hegeman, 1974, 1979). The sampling locality is the NIOZ Marsdiep pier (A in Fig. 1), where weekly (more frequently during spring blooms, less frequently in winter) samples were taken with a bucket during high tide. Annual average values are based on monthly averages to compensate for this unequal sampling over the year. For chlorophyll-*a* measurements, 0.5 to 1 l water was filtered over $MgCO_3$ coated glass fibre filters (Whatmann GF/C); acetone extracts were measured following Lorenzen (1967). Phytoplankton was counted in Lugol preserved samples with the Utermöhl (1958) method using a Zeiss inverted microscope. Most larger species were identified to the species level, but many of the smaller flagellates had to be aggregated because of identification problems.

Results

Species composition

In Table 1 all dominant diatoms (occurring at least in one month between 1980 and 1990 in numbers exceeding 100 per ml) and *Phaeocystis* sp. have been enumerated. Species dominance and composition varies markedly from year to year. Only *Phaeocystis*, *Thalassiosira* sp., and *Skeletonema costatum* occurred in numbers >100 ml^{-1} in all years, albeit with marked differences in the number of months per year.

From Table 1 it is possible to list a Marsdiep phytoplankton top-ten. Below I give (in parentheses) the number of months the species was found (at least once) in numbers >100 ml^{-1}. *Phaeocystis* (81, most probably *P. globosa*, see Cadée, 1991) is numerically the most important plankton species, its period of occurrence in numbers >100 cells ml^{-1} increased from 4–6 months in the early 1980s to 10–11 months in the late 1980s. The colonial *Chaeotoceros socialis* (51) and *Skeletonema costatum* (49) follow in importance. *Thalassiosira* sp. (25), *Plagiogramma brockmanni* (24), *Asterionella glacialis* (24), and *Rhizosolenia delicatula* (24) are numbers 4 to 7; *Nitzschia seriata* (17) is number 8. *Melosira granulata* (16) number 9, is a freshwater species sluiced into the Wadden Sea from the Lake IJssel, particularly in winter and early spring. *Melosira sulcata* (10) and *Leptocylindrus minimus* (10) share 10th place. The latter species rapidly increases: it occurred for prolonged periods in numbers >100 ml^{-1} only in 1989 and 1990 and particularly in summer and autumn (Table 1).

Seasonal variation of Marsdiep phytoplankton

The seasonal variation for 1989 is given here as an example (Fig. 2). A diatom spring peak occurred early in April, followed by a *Phaeocystis* peak at the end of April. This is the usual sequence, the difference in time of the two peaks being on average two to three weeks (Cadée, 1990b; Cadée and Hegeman, 1991). The explanation for this peak sequence is that diatoms first deplete Si, leaving P and N in considerable amounts for algae not using Si (Bennekom *et al.*, 1975). The large second peak of diatoms in late summer, mainly consisting of the small *Leptocylindrus minimus*, was not observed in earlier years, but in 1990 it was observed again. Other non-diatom plankton reached highest numbers after the *Phaeocystis* peak. The yearly occurring peak in chlorophyll *a* coincides with the *Phaeocystis* spring peak; the diatom peaks in spring and late summer cause less marked chlorophyll *a* peaks.

Year-to-year changes in duration of blooming periods

Blooming periods are defined here as periods with >1000 cells ml^{-1}. The duration of these periods can easily be read from annual curves giving variation in cell numbers over the year. Figure 2 gives an example of such a curve for 1989; curves for earlier years were published in Cadée (1990b). To compare the different years, the length of this blooming period is in my experience the best parameter, better than, for instance, average or maximum cell numbers, values which are severely influenced by missing or sampling the peak values. For the period 1980 to 1990 it is difficult to discern trends, but if we include older data (Fig. 3) it appears that blooming periods of flagellates and *Phaeocystis* increased since the early 1970s. Diatom blooming periods varied considerably, increasing from short dura-

Table 1. Dominant phytoplankton algae (diatoms and *Phaeocystis*) for the period 1980–1990 occurring at least once in densities >100 per ml in the months indicated as numbers. *Asterionella glacialis* is synonym for *A. japonica*; *Chaetoceros socialis* includes *C. radians* and probably some other small *Chaetoceros* sp.; *Chaetoceros* sp. includes all larger species.

Species	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989	1990
Skeletonema costatum	3456	3456	34568	3456789	3479	346	34567	3456789	34568	3456	3
Chaetoceros socialis	4789		789	46789	4789	345678910	459	345689	345689	456789	4567912
Chaetoceros sp.	8					9	8	58		7	7
Rhizosolenia delicatula	4568		56789	456	6	45	4568	6	5	510	8
R. shrubsolei						78		8	78	7	
R. hebetata				6	7		8		5		
R. setigera					6						
Biddulphia aurita	3			4	34	4	4	5			4
Ditylum brightwellii									4		4
Cerataulina bergonii	6		6	8					5		5
Leptocylindrus danicus				8						8	
L. minimus										78910	1357810
Melosira sulcata			349	47		512	34				3
Thalassiosira sp.	34	4	345	4567	3	34611	3467	347	6	4	4
Asterionella glacialis	45	45	45	4568	45	456	45	346	45		45
A. kariana	3	4	2345			3		3		4	
Eucampia zoodiacus							6				
Thallassionema nitzschioides		4	4	4		3	3				4
Nitzschia closterium	4										
N. longissima				4	4		4				8
N. seriata	567	56	4	67	56	456		6	4	45	
N. delicatissima									5		
Plagiogramma brockmannii	34	45	345	4579	45	2346	34	3456			4
P. vanheurckii			5	6							
Cymatosira belgica			58		9		910				
Phaeocystis	4–9	4–7	3–8	3–8	4–7	2–10	4568910	4–11	2–12	3–12	2–11
(Freshwater species)											
Diatoma elongatum	23						4				
Melosira granulata	12312	12	3		13	23	24		14		Dec

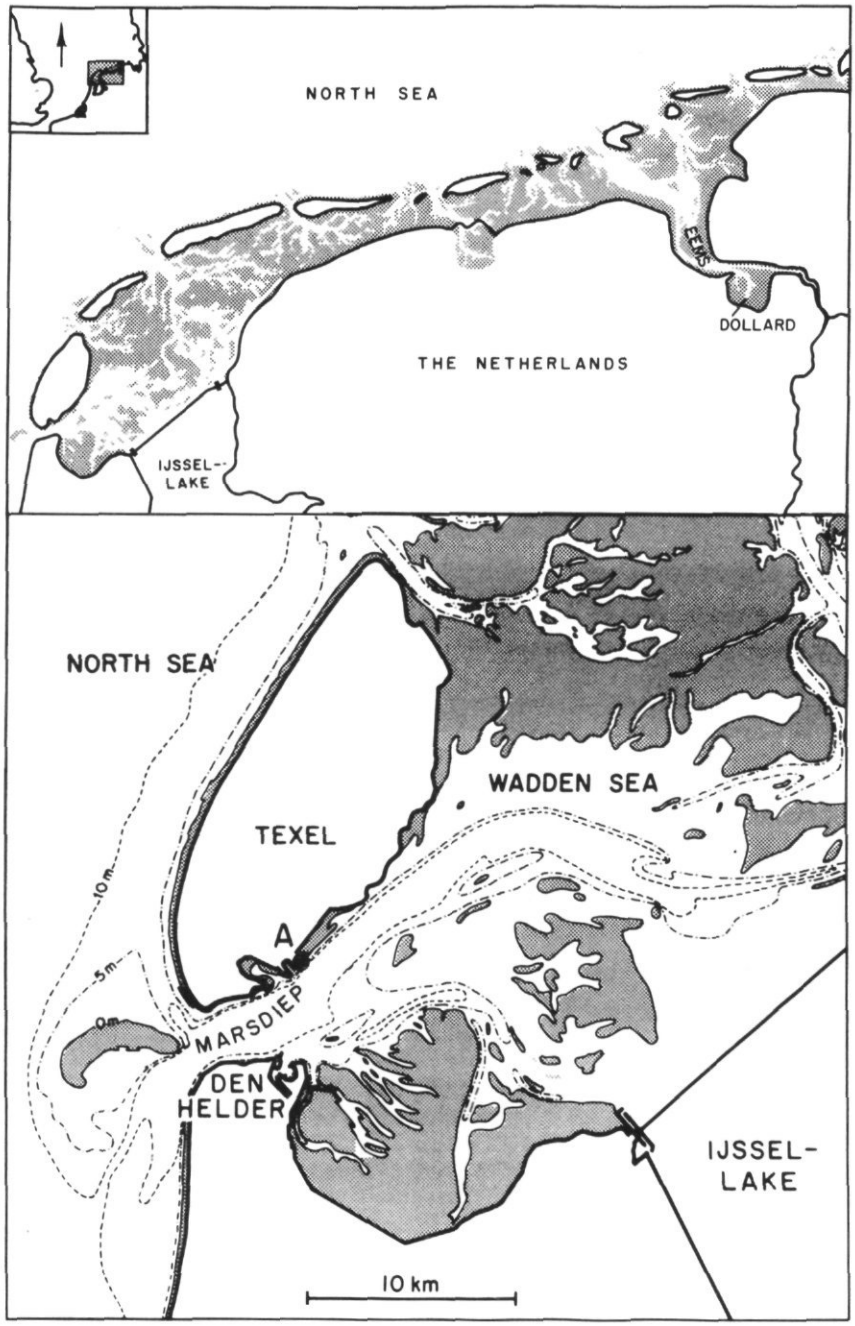

Figure 1. Location of the Marsdiep sampling station (A).

tion in the early 1970s to long blooming periods in the mid-1980s, but in the late 1980s they decreased again to a length similar to that in the late 1960s (Fig. 3a).

Year-to-year changes in chlorophyll *a*

Annual average chlorophyll *a* (Fig. 4) showed variations in the period 1980–1990, and again, adding older data, we can discern an increase in the late 1970s up to the mid-1980s comparable to the increase in the duration of blooming periods of non-diatoms. After the mid-1980s the average chlorophyll *a* content drops again, but remains above the values for the early 1970s.

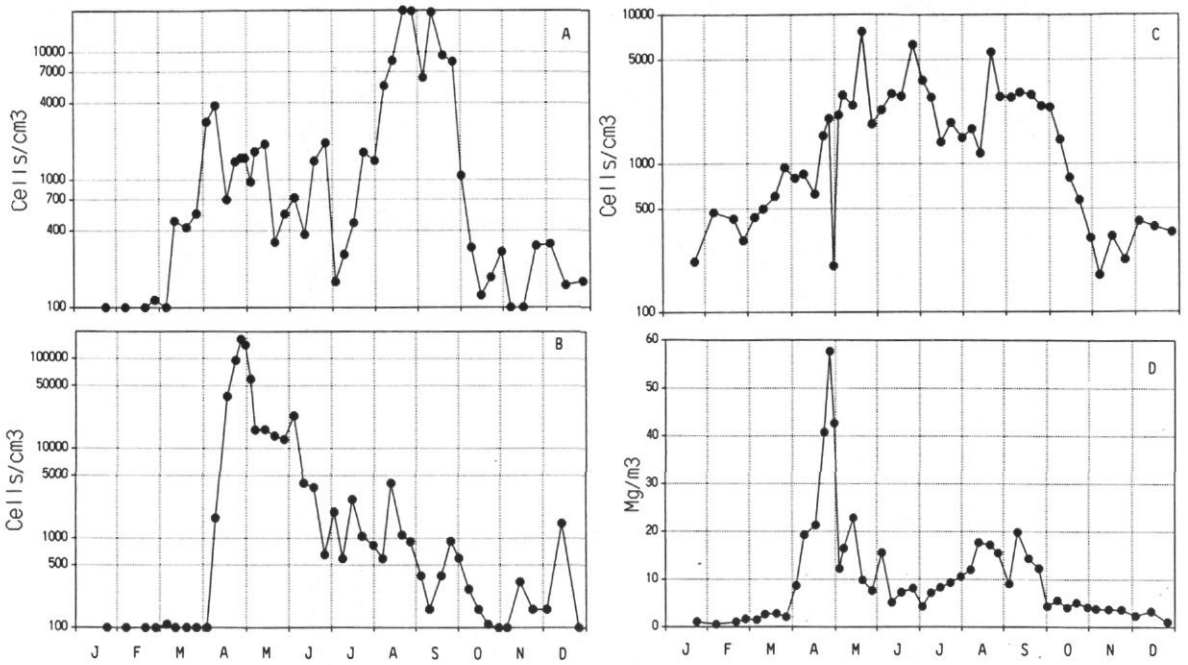

Figure 2. Seasonal phytoplankton cycle at the Marsdiep sampling station in 1989. A = diatoms, B = *Phaeocystis*, C = other non-diatoms (note differences in vertical scale), D = chlorophyll *a*.

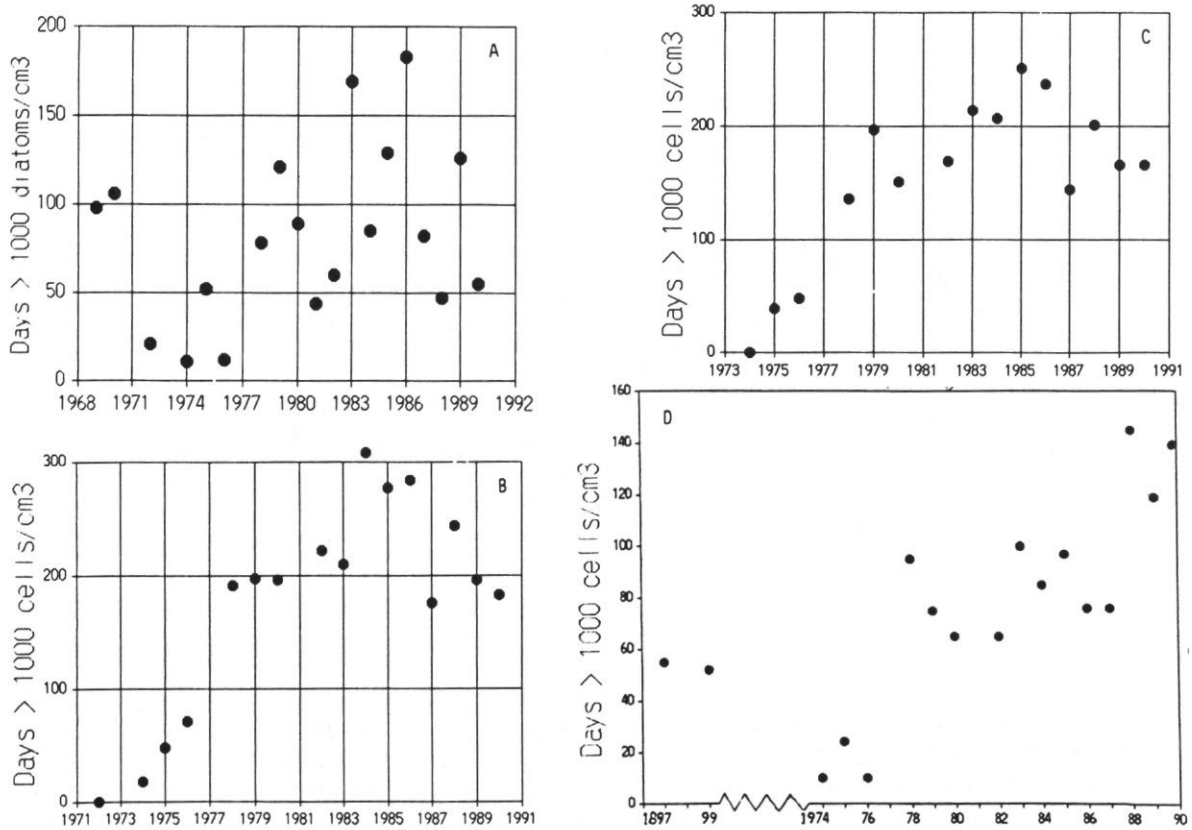

Figure 3. Length of "algal blooming periods" (days with cell numbers exceeding 1000 cells · ml^{-1}) in the Marsdiep for the different years. A = diatoms, B = all non-diatoms, C = non-diatoms minus *Phaeocystis*, and D = only *Phaeocystis* and including historical data of Cleve (1898–1900).

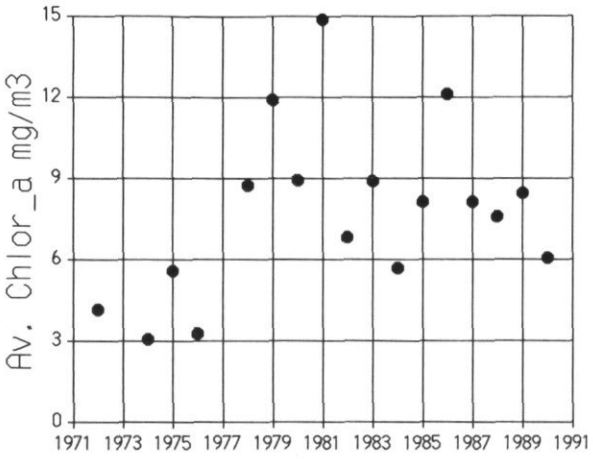

Figure 4. Year-to-year variation in annual phytoplankton biomass in the Marsdiep measured as annual average chlorophyll *a*.

Discussion

Eutrophication

Changes observed in phytoplankton: increase up to the mid-1980s in blooming periods, in biomass (chlorophyll *a*) and in primary production (Cadée, 1986b, 1992), have been attributed to eutrophication (Cadée, 1986b; Beukema and Cadée, 1988; de Jonge, 1990). The higher inputs of nutrients from the rivers, particularly the Rhine, are held responsible. However, increase in nutrient concentrations in the Rhine started much earlier, around 1950 (van Bennekom *et al.*, 1975; van Bennekom and Salomons, 1980). Part of the Rhine input enters the Wadden Sea via Lake IJssel, and van der Veer *et al.* (1989) have argued that the late response of Marsdiep phytoplankton to eutrophication (only in the 1970s, whereas rapid increase in nutrients from the Rhine as measured near Lobith started 20 years earlier and stabilized in the 1980s) was due to storage of nutrients (P) in the bottom sediments of Lake IJssel. When these sediments were saturated they became a source themselves for nutrients, which resulted in higher nutrient loadings from Lake IJssel to the Wadden Sea in the late 1970s (Raaphorst and van der Veer, 1990). Riegman *et al.* (1992a,b) point to the role of the building of the Houtribdam in 1975, disconnecting the southern part of Lake Ijssel where muds with high binding capacity for P occurred and thus leading to increased nutrient discharges to the Wadden Sea after 1975. Van der Veer (1990) hypothesized that the introduction of an extensive mussel culture in the western Wadden Sea in 1950–1960 (whereas since 1960 the area used for mussel culture remained constant) might have counteracted earlier effects of eutrophication on phytoplankton. Cadée and Hegeman (1974) actually observed lowering of phytoplankton biomass and primary production in summer above a mussel culture area in the western Wadden Sea. In conclusion, a number of causes might explain a retarded influence of eutrophication on the Wadden Sea/Marsdiep area.

The eutrophication by (large) rivers of the ocean in front of the rivermouth leading to areas in the plume of higher phytoplankton biomass and primary production has been found repeatedly (e.g. Hobson, 1966; Cadée, 1975, 1978; Revelante and Gilmartin, 1976; Sklar and Turner, 1981; Ning *et al.*, 1988; Lohrenz *et al.*, 1990). Szekielda (1982, 1983) and Szekielda and McGinnis (1991) note that the area of eutrophication is of vital importance for the coastal ecosystem and may play a more important role than thought in the past; its presence and spatial variability can be studied best by using satellite observations. This eutrophication may lead to large algal blooms and subsequent anoxia with mass mortalities of benthos and fish if river inputs of nutrients are increased too much due to the activity of men. The Roumanian coast along the Black Sea influenced by the Danube (Gomoiu, 1983, 1987) and the northern Adriatic influenced by the River Po (Marchetti *et al.*, 1988, 1989) are good examples.

The Rhine influence

Reid *et al.* (1983, 1990) stress the importance of river influence on coastal North Sea phytoplankton. Increased river discharges add extra nutrients and may stabilize the water column, thus reducing vertical mixing and helping phytoplankton to stay in the euphotic zone. Gieskes and Schaub (1990) and Schaub and Gieskes (1991) observed a correlation between Rhine discharge in spring and summer and phytoplankton biomass in the Dutch coastal zone in the same periods. Nutrient loads (kg/sec) in the Rhine entering The Netherlands have stabilized in the 1980s (nitrogen), and even a decrease in the discharge of total dissolved P occurred due to measures like better wastewater treatment in Germany, which might influence fish production negatively (Boddeke and Hagel, 1991).

Increased water discharge gives lower total P concentrations (mg l^{-1}) at Lobith (Fig. 5a, $r = 0.85$, $p = <0.0005$; see also Gieskes and Schaub, 1990: fig. 6) due to dilution, and there is no significant increase in P discharge (kg sec^{-1}) with increased water discharge (Fig. 5b, $r = -0.07$). Gieskes and Schaub (1990) argue that increase in rainfall enhances nutrient leaching from the nutrient-saturated soils in The Netherlands and higher current velocities during high river discharge enhance exchange of nutrients from interstitial water in the Dutch part of the river Rhine. As a result, more nutrients should be added to the Dutch part of the catchment area of the Rhine during high water discharges and consequently more nutrients are carried to the North Sea. Nitrogen data not used by Gieskes and

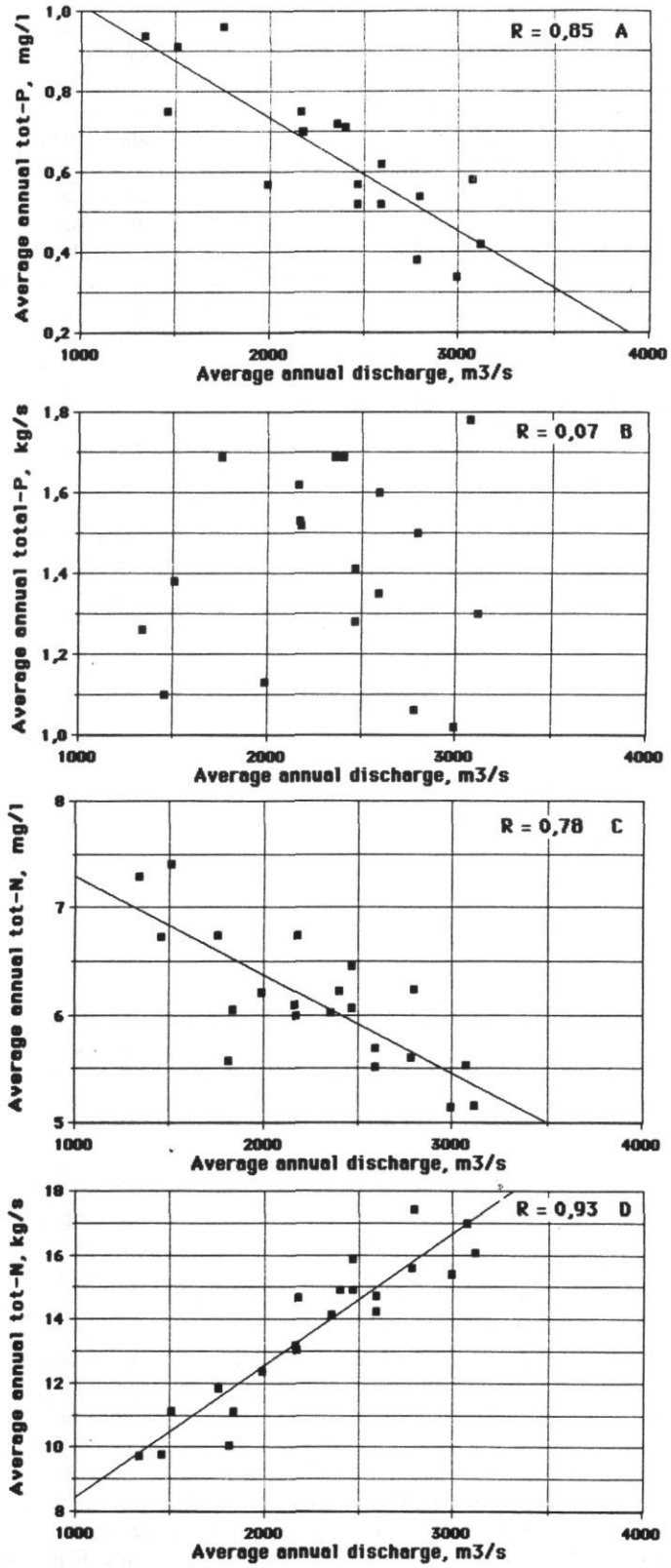

Figure 5. Comparison of average annual Rhine discharge at Lobith (1970 to 1988, data RWS) with: A – average annual total P concentration ($mg \cdot l^{-1}$), B – average annual total P discharge ($kg \cdot s^{-1}$), C – average annual total N concentration ($mg \cdot l^{-1}$), D – average annual total N discharge ($kg \cdot s^{-1}$).

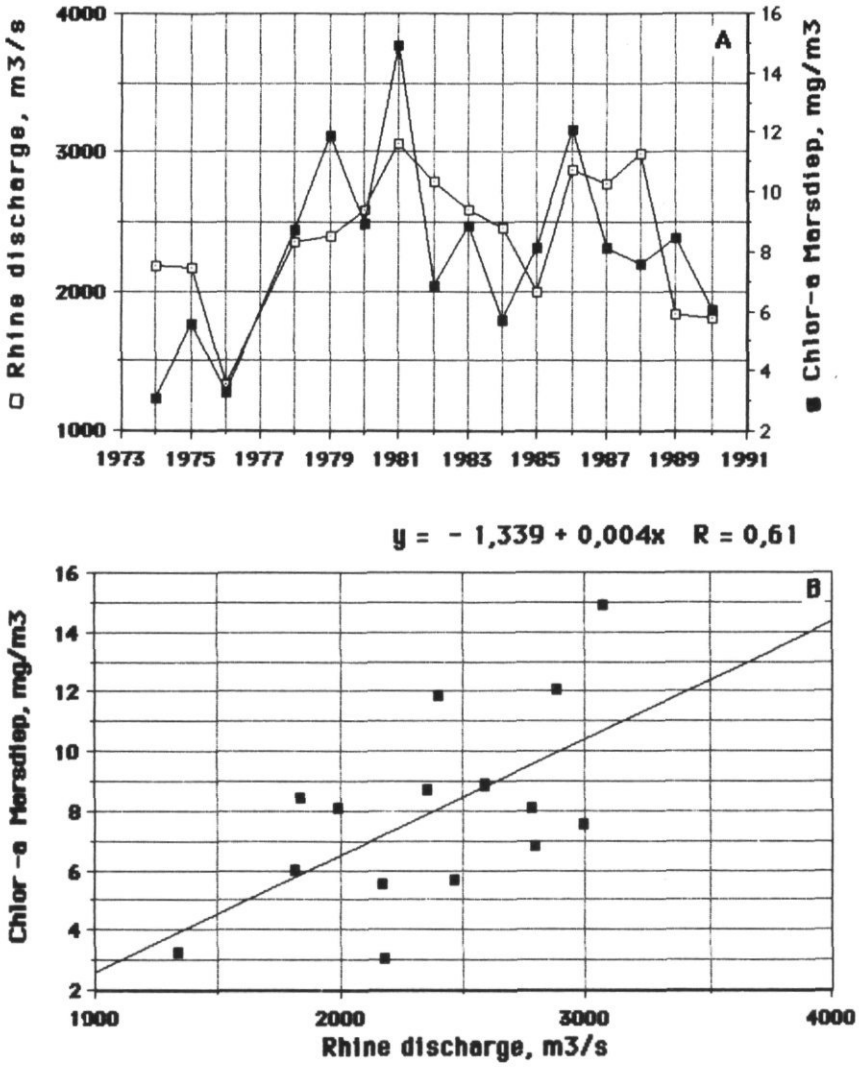

Figure 6. Comparison of annual average chlorophyll *a* of the Marsdiep and Rhine runoff at Lobith. A = year-to-year variation, B = correlation.

Schaub illustrate this process also for the German part of the catchment area of the Rhine: there is some dilution of total-N with higher river discharge (Fig. 5c), but there is a highly significant positive correlation between N-load (kg sec^{-1}) and Rhine discharge (Fig. 5d). In order to estimate the total input of nutrients by the Rhine we have to add to those discharged at Lobith those contributed in The Netherlands. Certainly, years with high rainfall (and thus high Rhine discharge) will also be years with high N-leaching from Dutch overfertilized agricultural soils. However, because of the spreading of Rhine water over different tributaries in The Netherlands this is more difficult to evaluate.

In order to test the hypothesis of Gieskes and Schaub (1990) of a causal positive correlation between Rhine discharge and Dutch coastal phytoplankton, which was based on a data set limited to the period 1977–1982, I compared quarterly and annual chlorophyll *a* values for the Marsdiep with Rhine discharge data for the period 1974–1990. A good correlation was found between annual discharge of the Rhine (Lobith) and the annual average chlorophyll *a* content of the Marsdiep (Figure 6, $r = 0.61$, $p = 0.01$). The relatively low chlorophyll *a* data for the early 1970s can be explained by low Rhine discharge values. The highest annual chlorophyll *a* value (1981) corresponds with the highest Rhine discharge value.

Average values for Rhine discharge and Marsdiep chlorophyll *a* for the separate quarters of the year did not show a correlation. Gieskes and Schaub (1990) found positive correlations during the 2nd and 3rd quarters, i.e. during the growing season of phyto-

plankton. This indicates a direct influence of Rhine discharge of the coastal area near the Rhine mouth, also observed in 1977 by Reid *et al.* (1983). The Marsdiep, farther away from the Rhine mouth, is influenced less directly. Water transport from the Rhine mouth to the Marsdiep takes one month on average (van Bennekom *et al.*, 1974: fig. 5). Moreover, the Marsdiep area is also influenced by Rhine discharges via the river IJssel and Lake IJssel; this is a longer way with a flushing time of Lake IJssel in the order of 0.5 year (van Bennekom *et al.*, 1974: fig. 5). Van der Veer *et al.* (1989), de Jonge (1990), and Riegman *et al.* (1991, 1992) deal in more detail with the importance of Lake Ijssel hydrography and dike construction on the amounts and ratios of nutrients introduced into the western Wadden Sea and influencing also the Marsdiep area.

Conclusion

Marsdiep data corroborate observations by Gieskes and Schaub (1990); high Rhine discharges are correlated with higher phytoplankton biomass in the Dutch coastal zone. An increase in discharge leads to an increase in nutrient discharge, particularly of N. The increased nutrient inputs leading to these higher phytoplankton biomasses must partly originate from the Dutch part of the catchment area of the Rhine. For the Marsdiep area they will arrive via Lake IJssel, too.

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The role of environmental changes in an unusual coastal plankton bloom

Mário E. C. Vieira, Elizabeth M. Cosper, and Edward J. Beltrami

Vieira, M. E. C., Cosper, E. M., and Beltrami, E. J. 1992. The role of environmental changes in an unusual coastal plankton bloom. – ICES mar. Sci. Symp., 195: 223–231.

Certain embayments on the northeast coast of the United States experienced previously undocumented microalgal blooms called "brown tides" during the summer of 1985. A small (2–3 μm), previously unidentified chrysophyte species, *Aureococcus anophagefferens*, was found to be responsible for these blooms. The blooms recurred in several bays during the summer of 1986, subsequently waning in 1987 and 1988, and finally disappearing in 1989. Investigations of causes of these novel blooms have indicated that perturbations in climate were in part responsible. Specifically, the onset of drought conditions and variations in wind regimes resulted in hydrological changes in the affected areas. Comparison of significant environmental variables between bloom and non-bloom areas throughout the 1980s has elucidated important regional and local factors in bloom initiation and maintenance. A dynamical systems model has been utilized to determine the threshold levels of key variables and to assist in understanding the underlying mechanisms which lead to the onset of such unpredictable bloom events.

Mário E. C. Vieira: United States Naval Academy, Oceanography Department, Annapolis, Maryland 21402-5026, USA; Elizabeth M. Cosper: Marine Sciences Research Center, State University of New York, Stony Brook, New York 11794, USA; Edward J. Beltrami: Department of Applied Mathematics, State University of New York, Stony Brook, New York 11794, USA.

Brown tide blooms

Several coastal embayments along the northeast coast of the USA during the mid-1980s have experienced novel microalgal blooms for which there is no previous record. These monospecific blooms were popularly called the "brown tide" due to the resulting water colour. In the early summer of 1985 the first appearance of the "brown tide" occurred over a wide geographic range along the coast in non-contiguous bodies of water: Narragansett Bay, Rhode island, and Long Island embayments in New York as well as Barnegat bay in New Jersey (Fig. 1) (Cosper *et al.*, 1987; Sieburth *et al.*, 1988; Nuzzi and Waters, 1989; Olsen, 1989; Sieburth and Johnson, 1989; Smayda and Villareal, 1989). The extent of the blooms was restricted to these coastal bay systems; blooms did not appear to follow a pattern of spreading from one bay system to the next. This suggests that the environmental factors contributing to these "brown tide" blooms were not just localized to specific conditions in a bay system but probably were more regional, e.g. involving meteorologically induced changes. The blooms on Long Island markedly reduced the extent of eelgrass (*Zostera marina*) beds because of increased light attenuation, and decimated populations of commercially valuable bay scallops (*Argopecten irradians irradians*) since the scallops were unable to graze adequately and starved to death (Cosper *et al.*, 1987; Bricelj and Kuenstner, 1989; Dennison *et al.*, 1989). Similarly, in Narragansett Bay the mussels were unable to feed and populations were severely reduced (Tracey *et al.*, 1989).

In 1986 the blooms recurred throughout the summer months in the same Long Island embayments as previously. In Long Island and in Barnegat Bay, NJ during the summers of 1987, 1988, 1989 the "brown tide" blooms returned only in diminishing levels (Fig. 2) (Cosper *et al.*, 1989a; 1990; Nuzzi and Waters, 1989; Olsen, 1989). During the summer of 1990, just as everyone was convinced that they were waning away, very brief but substantial outbreaks of the bloom were recorded in certain Long Island bay areas by the monitoring program of Suffolk County (Nuzzi, 1990), indicating that the potential for these blooms to recur still exists. Since 1985 "brown tide" blooms have not returned to Narragansett Bay (Sieburth *et al.*, 1988; Smayda and Villareal, 1989).

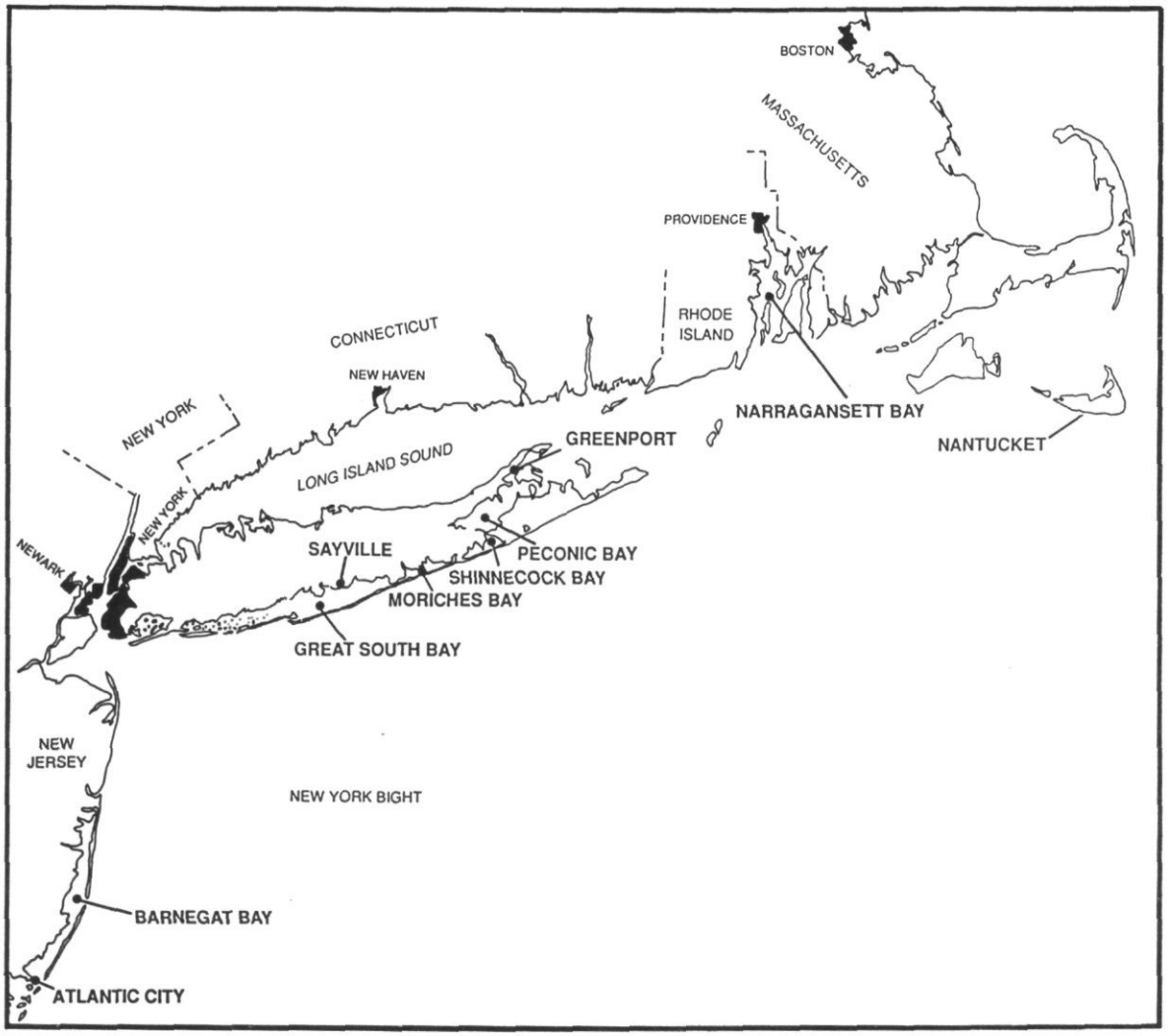

Figure 1. Regional map of northeast coastal areas including bays affected by the brown tide.

The unique aspect of these blooms was the dominance of a single very small (2–3 μm in cell diameter) microalgal species in terms of cell number (greater than millions of cell ml^{-1}) which contributed greater than 80% of total cellular phytoplankton volume throughout most of the bloom period during the summer months (Cosper *et al.*, 1987, 1989a). Usually during the summer months "small form" species are dominant, but as a diverse group of species with constantly varying successional patterns. During the blooms phytoplankton biomass, as indicated by chlorophyll *a* levels, was not particularly elevated for Long Island bays, in comparison to other years (Fig. 3), since concentrations reached levels less than 30 $\mu g\, l^{-1}$ during the height of the bloom in 1986 and chlorophyll *a* was concentrated in the smaller (less than 5 μm) fraction consistent with the cellular size (Bruno *et al.*, 1983; Lively *et al.*, 1983; Cosper *et al.*, 1987, 1989a). Only during June of 1989 when the "brown tide" bloom was absent were extremely elevated chlorophyll concentrations measured at 240 $\mu g\, l^{-1}$ which were associated with a dinoflagellate bloom. Primary productivity levels were high, but also were no different from pre-bloom years and estimates of phytoplankton carbon turnover during summer months were rapid, of the order of hours except during the dinoflagellate bloom in June 1989 (Fig. 4) (Cosper *et al.*, 1989a). Changes in inorganic nutrient levels, as nitrate, nitrite, phosphate and ammonium, were no different from pre-bloom years (Hardy, 1976; Bruno *et al.*, 1983; Lively *et al.*, 1983). Variations in inorganic macronutrients were not correlated with variations in the productivity of the "brown tide" and there is no evidence to support increased macronutrient loading, such as might come from sewage treatment plants, as a cause of the blooms (Cosper *et al.*, 1989a, b). These findings are consistent with similar studies in Narragansett Bay,

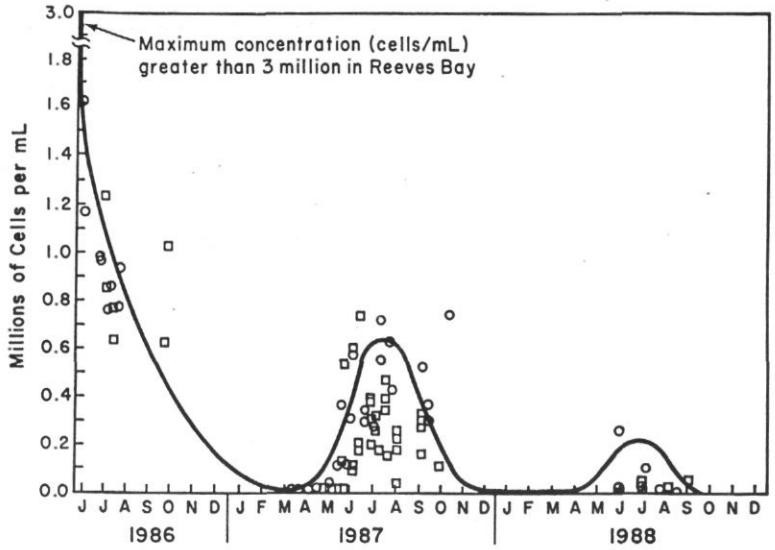

Figure 2. Brown tide cell concentrations for sites sampled in Peconic (□) and Great South Bays (○): 1986, 1987, and 1988.

Rhode Island (Smayda and Villareal, 1989; Keller and Rice, 1989).

The "brown tide" alga, *Aureococcus anophagefferens* (Sieburth *et al.*, 1988), is a newly described species not previously known to cause blooms, thus environmental conditions contributory to the blooming could in part relate to new anthropogenic influences in these bays. Physiological studies have identified certain micronutrients as conducive to the growth of this species, such as specific chelators (which have replaced phosphates in

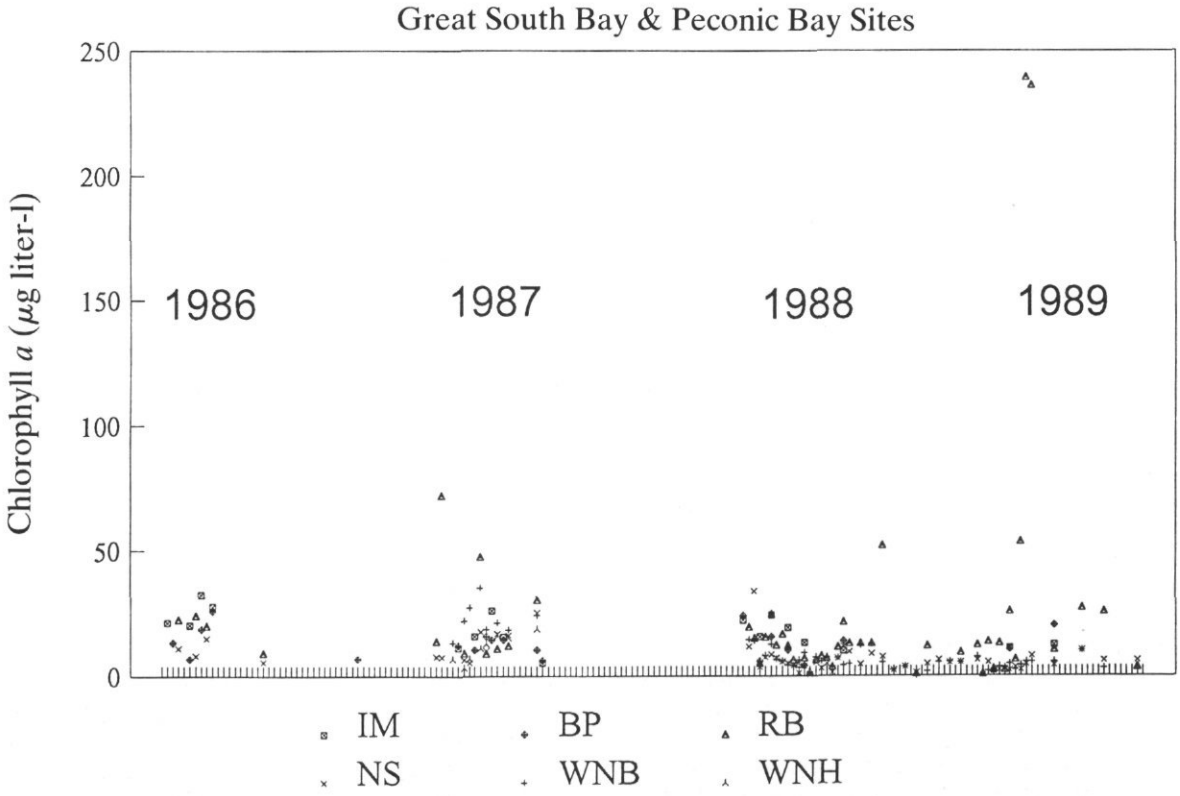

Figure 3. Chlorophyll *a* concentrations for sites sampled in Long Island bays from 1986 to 1989.

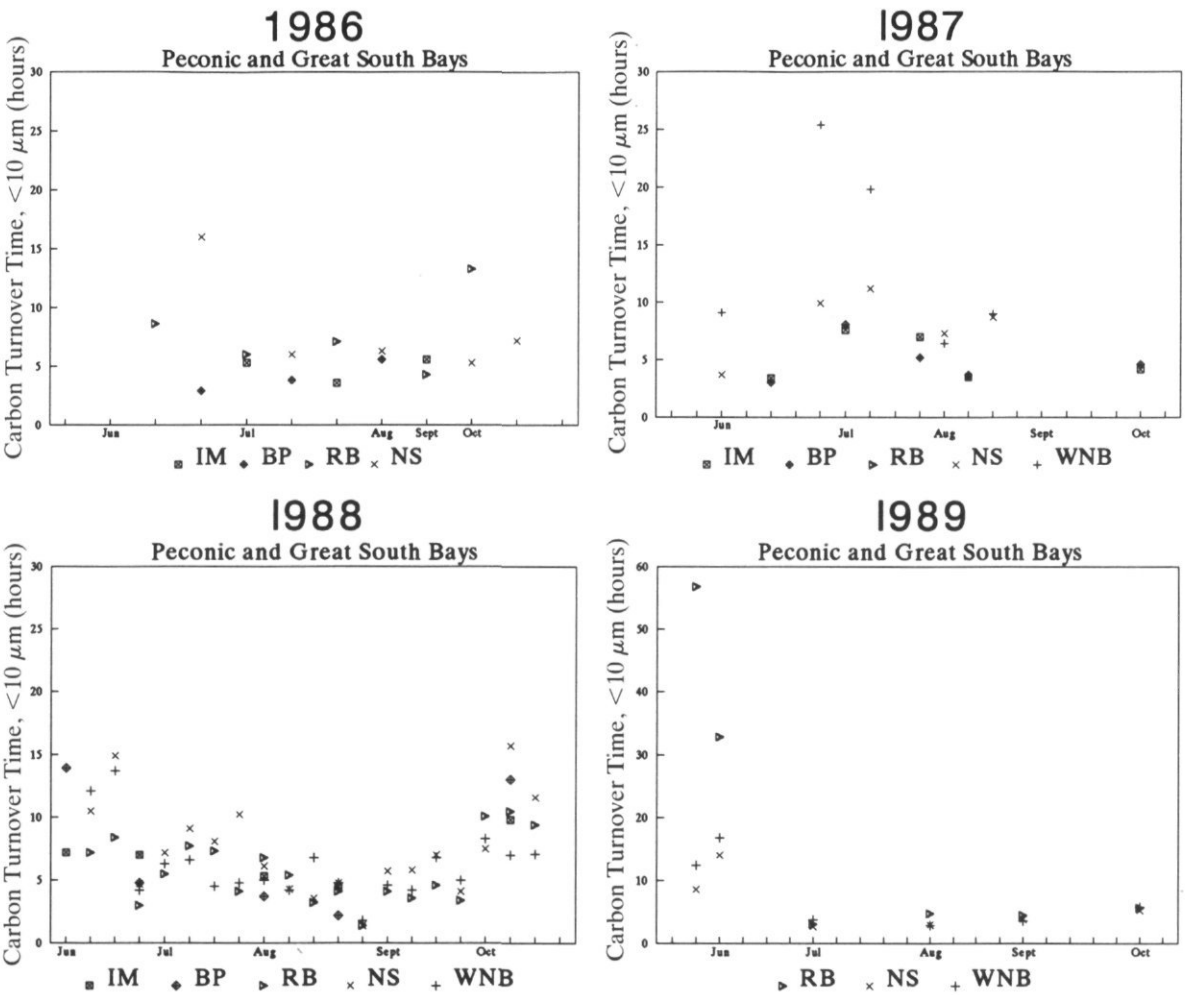

Figure 4. Carbon turnover times for sites sampled in Long Island Bays from 1986 to 1989.

detergents), selenium and iron, and organic phosphates in nitrogen sources (Cosper *et al.*, 1990). In addition, rainfall records since 1949 indicated that the first year of the bloom, 1985, started a drought which persisted into 1988 (Fig. 5). In 1989 rainfall levels exceeded all previous records. The drought conditions served to elevate salinities in the bays to a level of 30 parts per thousand conducive to the growth of *A. anophagefferens*, along with pulses of rain which delivered specific nutrients to the bay waters (Cosper *et al.*, 1990). Since this species does not appear to be very palatable to many organisms, selective grazing pressures against the "brown tide" species during the early bloom phase could have further allowed for the development of large populations (Caron *et al.*, 1989). The possibility that pesticides in certain fertilizers could have killed the microconsumers in the water column, thus allowing for the proliferation of the "brown tide" species, has also been suggested (Sieburth, 1989).

No single environmental factor caused these blooms and research still continues to evaluate the possible role of man-made influences. Since these bays are naturally restricted in terms of flushing by coastal waters, the possibility of increased residence time of water and physical stability in the bays contributing to these blooms, as has previously been invoked for other blooms (Kierstead and Slobodkin, 1953), was investigated.

Hydrodynamics

A phenomenon with the magnitude of the *Aureococcus* bloom of 1985 is bound to be the result of a very complex chain of events with a number of physical, biological, and chemical factors occurring at the right place and at the right time. In the search for a plausible physical contribution to this phenomenological continuum, we

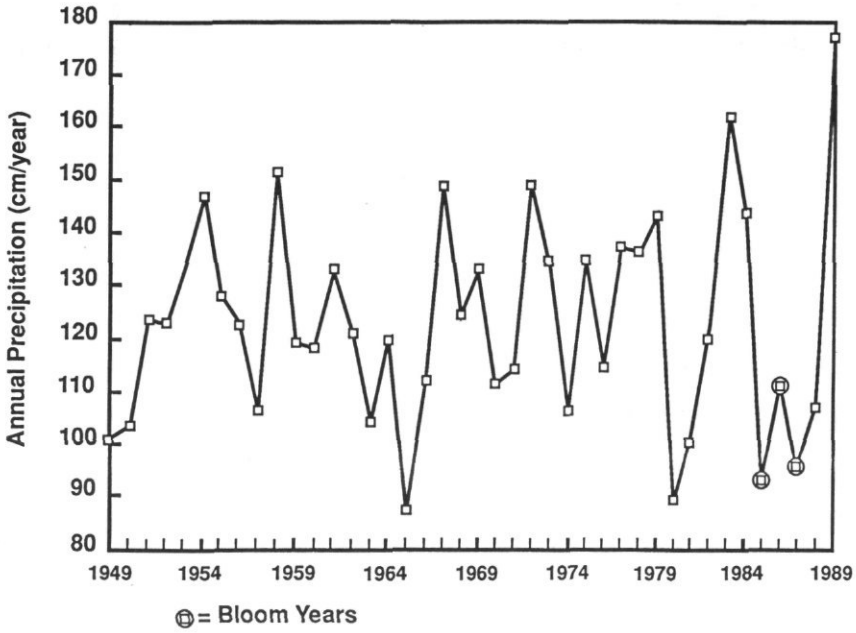

Figure 5. Total monthly precipitation measured at Brookhaven National Laboratory, Long Island, from 1949 to 1989.

were reminded of what Legendre and Demers (1984) designated as "dynamic biological oceanography" in recognition of the role of hydrodynamics as the driving force of aquatic ecosystems. The incidence of the bloom at the regional level and not as an isolated event was interpreted as an indication of possible meteorological forcing acting through an alteration of the flushing regime of the affected embayments. Our suspicions were reinforced by recent findings of a relationship between the bloom of a microorganism and estuarine flushing (Garcon *et al.*, 1986) as well as about the influence of wind forcing on the subtidal residence times of estuarine waters (Goodrich, 1988).

In the course of a first look at the phenomenon, Vieira (1989) hypothesized that hydrodynamic changes in the waters subjected to the "brown tide" bloom were determined by marked meteorological perturbations. Specifically, he suspected that lowered levels of energy in the winds throughout the region had led to increased residence times of the impacted waters; this might in turn mean longer persistence and availability of the biochemical forcing responsible for the rapid growth of the picoplankter. This result was aggravated by conditions of moderate to severe drought in the spring of 1985, which lowered the freshwater inflow rates and promoted slightly higher salinities (Cosper *et al.*, 1989b).

The present investigation was centered on the analysis of sea level records at seven locations throughout the area from 1 January 1980 to 30 June 1988. Three of these tidal stations (Nantucket, Montauk, and Atlantic City, Fig. 1) were on the open coast and represented oceanic conditions; the other four (West Sayville in Great South Bay, Coast Guard Station in Moriches Bay, Shinnecock Dock in Shinnecock Bay, and Greenport in Peconic Bay, Fig. 1) were inside the affected estuaries. It was assumed that the deterministic forcing of the embayments at tidal frequencies would remain independent of any seasonal or annual meteorological fluctuations; subsequent analysis of the data proved this assumption to be correct. The time series were filtered to remove the astronomical tidal components and higher frequencies; subsets of the resultant low passed records were generated for the winter (December to February) and spring (March through May) months. The objective was to examine the oscillations of the subtidal component of the sea level during the years preceding and following the bloom outbreak and to identify any trends or seasonal changes.

The power spectra of these low frequency winter and spring sea level records were analysed. No evidence was found for any pattern of change in frequency bands with season. The spectral density functions were integrated and the total variance (representing energy) therein contained was computed. Figure 6 shows the distribution of the variance throughout the spring and winter seasons for the oceanic stations over the nine years of records. It is clear that in the spring of 1985 (the onset of the bloom was May 1985) the energetic content of the sea-surface sub-tidal oscillations was at an all-time minimum since the beginning of the decade. Furthermore, the decrease from winter to spring was the highest since 1982. Similar results were obtained for the estuarine stations. We interpret these observations of suddenly decreased sub-tidal sea level fluctuations to mean sig-

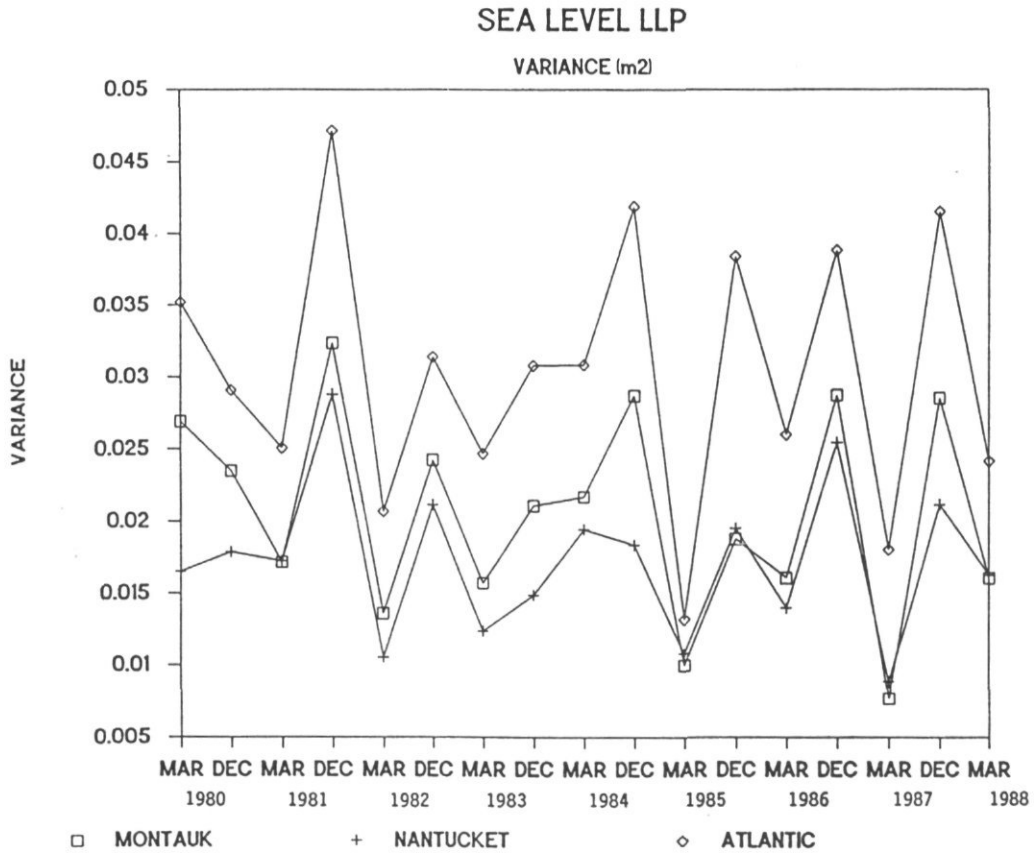

Figure 6. Seasonal fluctuations of sub-tidal sea level variance at the oceanic stations.

nificantly lowered sub-tidal volume fluxes throughout the embayments and thus reduced exchanges with the ocean, resulting in weaker flushing at these time scales.

The variance of the semi-diurnal flux and reflux of the waters to the estuaries due to the astronomical tide was also computed and found to be insensitive to the season throughout the nine years. It appears then that underlying the relatively quiescent tidal exchange there is a seasonally forced longer-term contribution, which hit a minimum in the spring of 1985 after having reached high values in the winter of 1984. Estimates of sub-tidal (long-term) residence times of the waters in the four estuaries were made, based on computed volume fluxes, the volume of the embayments and recirculation (mixing) coefficients extracted from salinity data. Figure 7 shows the relative increase of about 10% in sub-tidal residence time in the spring of 1985 for Great South Bay, Shinnecock, and Moriches. The absence of data for Greenport in 1985 prevents a conclusive analysis for the Peconic Bays; however, the existing data suggest a rising trend for spring residence times since 1983.

Wind records from Brookhaven National Laboratory (Long Island, NY), assumed representative of regional conditions, were examined. It was found (Fig. 8) that the variance of the East–West component of the wind stress dropped to a minimum in the spring of 1985. Since this component has been associated with the sub-tidal forcing of Great South Bay (Wong and Wilson, 1984) we conclude that reduced levels of regional wind activity in the spring of 1985 depressed the sub-tidal oscillations of sea level inside the estuaries, resulting in increased sub-tidal residence times for those waters; the tidal contribution, in contrast, was not found to show a seasonal dependency. This hydrodynamic scenario may have contributed to the chain of events leading to the "brown tide" outbreak in the spring of 1985.

Modelling results

In an attempt to assess the impact of climate fluctuations on bloom dynamics, a simple mathematical model has been formulated that looks at the linkage between trophic variables and external forcing terms such as temperature, drought-induced salinity, and changing residence times. The model consists of differential equations that denote the change over time of concentrations of the picoplankter *Aureococcus* and of grazer activity on this species (Beltrami, 1989a,b, 1991). The

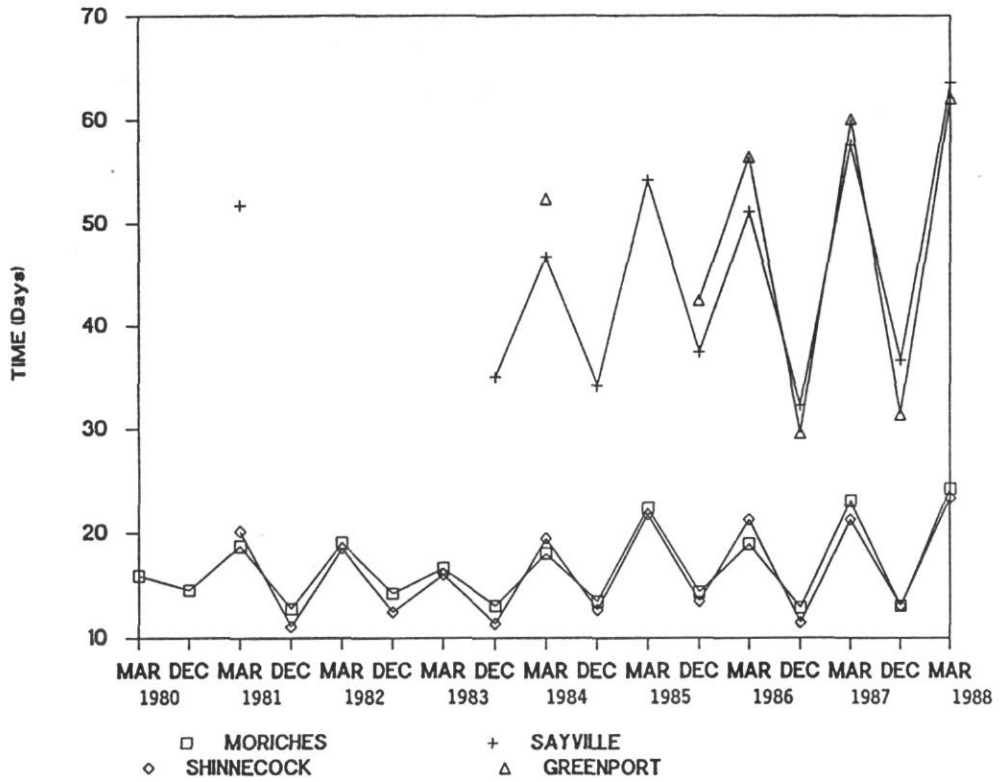

Figure 7. Seasonal fluctuations of sub-tidal residence times for the estuaries.

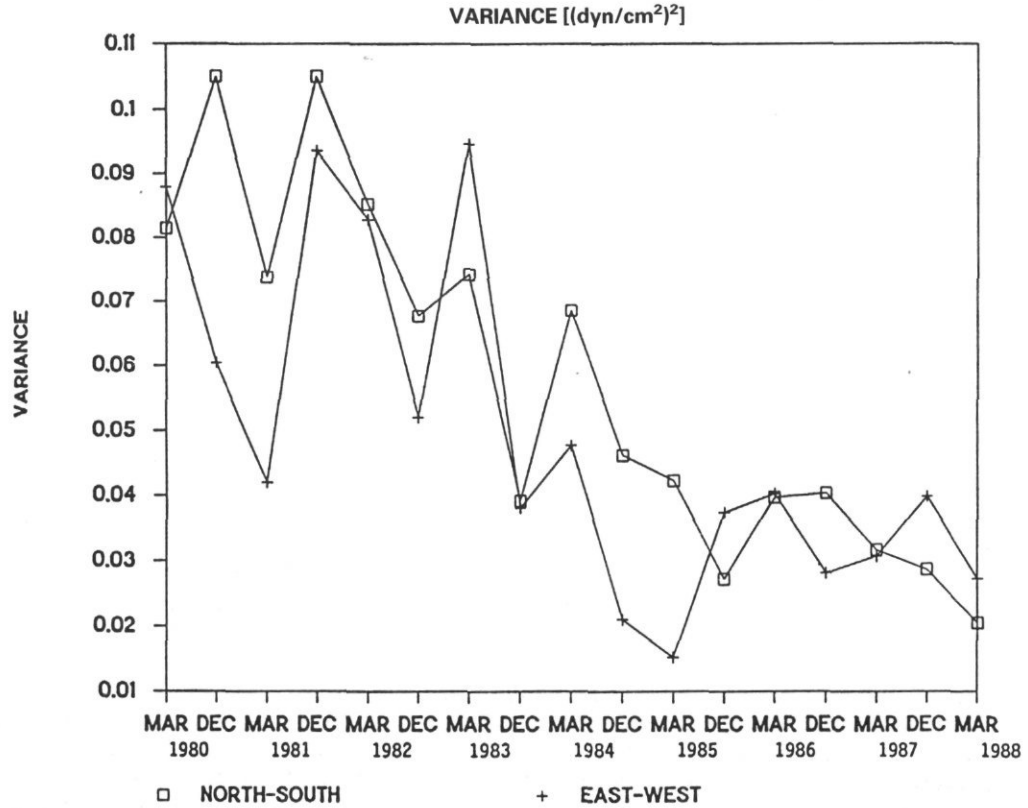

Figure 8. Seasonal fluctuations of low frequency wind stress variance at Brookhaven.

forcing terms consist of an annual temperature cycle and rainfall-induced changes in salinity, each of which separately limits the growth rate of *Aureococcus*. In addition, growth is limited by the residence time of certain micronutrients (Cosper *et al.*, 1989). Using a time scale of months, diurnal cycles are averaged out and the model focuses on multi-seasonal events.

One of the striking attributes of the model is that it exhibits thresholds in temperature, residence time, and salinity for the initiation of bloom episodes, below which no bloom manifests itself but beyond which there are sudden and dramatic surges in cell counts following a familiar "cusp catastrophe" scenario. Computations with the model show that the bloom is able to sustain itself for several years even if the meteorologically driven factors which induce the onset of a bloom are allowed to abate somewhat. This is seen as a hysteresis effect in which a bloom persists long into the autumn months at lower temperature and salinity regimes and, second, by its ability to rekindle itself in subsequent years after the environmental factors have altered considerably (Fig. 9).

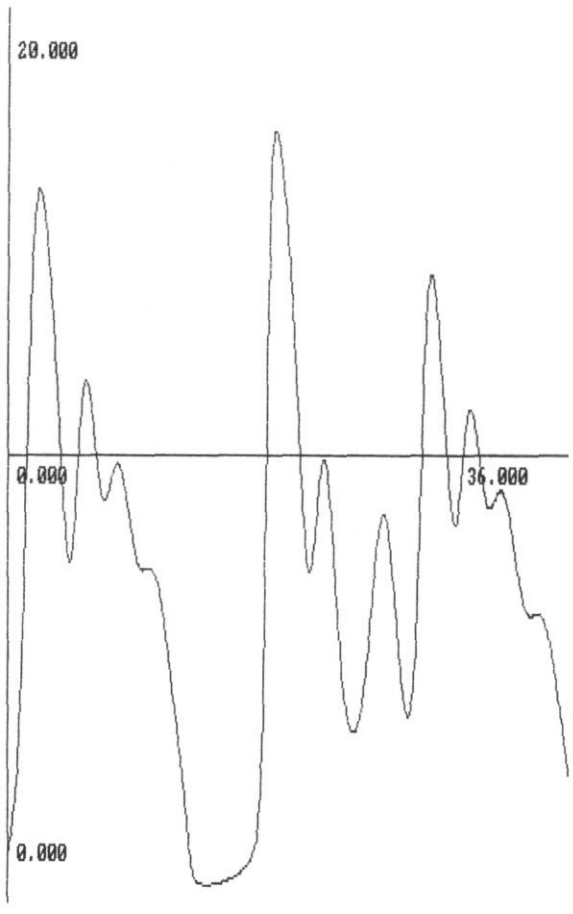

Figure 9. Model computed cell counts of *Aureococcus* in units of 5×10^7 cells l^{-1} as seen over a 36-month time span. The trophic model is here driven solely by periodic fluctuations in the annual temperature cycle. Observe that the bloom pattern does not repeat itself in the same way from year to year even though the climatic variability is chosen to be regular and predictable.

The model does not replicate exact cell counts and indeed it is doubtful that any model can accomplish this at present in view of the complexity of an actual estuarine system. Rather it mimics the underlying dynamics and it lends plausibility to the idea that once an episode of anomalous growth is triggered, the internal logic of the trophic chain may suffice to maintain unusual levels of activity. Moreover, the model suggests that erratic fluctuations in cell counts that one observes may be due as much to these trophic interactions as to the vagaries of climatic shifts.

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Recent changes in the summer plankton of the North Sea

R. R. Dickson, J. M. Colebrook, and E. Svendsen

Dickson, R. R., Colebrook, J. M., and Svendsen, E. 1992. Recent changes in the summer plankton of the North Sea. – ICES mar. Sci. Symp., 195: 232–242.

The results of the Continuous Plankton Recorder (CPR) survey are used to identify changes in the summer plankton of the North Sea during the 1980s. These changes are principally concerned with the *Ceratium* group of dinoflagellate species and are unprecedented in a CPR record of over three decades' duration. In 1987–1988, *Ceratium* species increased greatly in abundance in the north-central North Sea, apparently in response to a temporary change in the local climate. During these years, the highly stratified conditions that these species not only like but require were reinforced through the westward transport of fresh surface water from the Skagerrak and Norwegian coast under the stress of an anomalously strong easterly air stream.

R. R. Dickson: MAFF, Directorate of Fisheries Research, Fisheries Laboratory, Lowestoft NR33 0HT, UK; J. M. Colebrook: Plymouth Marine Laboratory, Plymouth PL1 3DH, UK; E. Svendsen: Institute of Marine Research, Nordnes, 5024-Bergen, Norway.

Introduction

It is now generally accepted that the ocean plankton may act as a sensitive indicator of climate change, with significant feedback effects on climate regulation. A recent Canadian report by Frank *et al.*, (1988) is typical of many reviews in anticipating that global warming in temperate and subarctic waters may act to favour components of the summer plankton, particularly the dinoflagellates: "Associated with a doubling of atmospheric CO_2 is increased vertical stratification of the water column due to higher freshwater discharge, higher temperatures and lower winds.... The net effect of increased water column stratification would shift the growth advantage from large diatoms to smaller (5–25 μm) dinoflagellates and microflagellates that ... are better able to maintain their vertical position in a stratified water column and utilize nutrients effectively at low concentrations.... Maintenance of an annual time series of the seasonal diatom:dinoflagellate ratio in continental shelf waters could provide an early warning signal for significant environmental changes."

However, summer is also the season when inorganic nutrient concentrations reach their minimum in coastal and shelf waters, when nutrient limitation of algal growth is likely to be at its most widespread, and therefore when changes in anthropogenic nutrient inputs are likely to have their maximal effect on the plankton. Not all of the planktonic changes associated with changing nutrient inputs concern eutrophication *per se*. The literature contains instances when nitrogen-fixing species have obtained a competitive advantage from the dwindling of riverine nutrient inputs through drought, and in the case of the 1988 *Chrysochromulina* bloom in the Skagerrak one significant feature of the bloom – its unusual toxicity – appears attributable to a phosphorus deficiency or, more properly to a low P:N ratio (Anon., 1990, p. 5). By and large, however, the expectation is that a summer increase in dinoflagellate bloom prevalence is also an expected result of anthropogenic eutrophication in coastal waters.

Thus the same relative shift in the community structure of the summer plankton is expected for both climatic and anthropogenic reasons and may well be part-driven by both in particular instances. After all, although the *Chrysochromulina* bloom already referred to may have owed its unusual toxicity to nutrient conditions, its essential characteristics were first modelled purely on the basis of the extreme water column stability that developed – for climatic reasons – during the winter/spring of 1987–1988. (Backhaus *et al.*, 1988).

In monitoring the health of our oceans it is important that the plankton monitoring system we employ should be capable of two things: first, it should be able to inform us *objectively* whether the prevalence, timing, or amplitude of a particular nuisance bloom has changed over the years; in a number of studies (e.g. Le Fevre, 1979; Mommaerts, 1986), increased reports of blooms have

simply been the result of more people looking, or the introduction of a more effective reporting system. Second, it should ideally be capable of distinguishing the global-climatic from the local-anthropogenic in the trends that we observe, since only when we know the relative impact of each will we know the extent and likely effectiveness of available countermeasures.

It is fortunate that in the case of the European shelf we already have multi-decade time series on plankton species-composition from the well-established Continuous Plankton Recorder (CPR) survey to identify trends in the plankton objectively, plus multi-decadal standard hydrographic section data from Norway to provide the basis to explain them.

A peak in *Ceratium* abundance, 1987–1988

The changes we describe in the summer plankton of the North Sea are principally – but not exclusively – concerned with a group of *Ceratium* species. Owing to their shape and relatively large size the Ceratia are collected quantitatively by the CPR, unlike many of the smaller dinoflagellate species.

In 1987–1988 the Ceratia increased greatly in abundance, on a scale unprecedented in a CPR record of over three decades' duration.

Figure 1 shows North Sea averages of the annual fluctuations in abundance (standardized to zero mean and unit variance) for three *Ceratium* species, *C. furca, C. fusus,* and *C. tripos*, over the period 1958–1988. The species themselves are depicted in Figure 2 (from Lebour 1925). Plainly, after three decades in which the annual fluctuations lay within 1 SD of the mean, (approximately), an increase took place in 1987–1988 which exceeded the long-term average by >+2 SD, without any apparent change in timing but with a parallel increase in the neritic copepods *Acartia* and *Temora*, and in the Cladocerans *Evadne* and *Podon* (Fig. 3).

Initially, because their time of peak occurrence coincided with the summer minimum in inorganic nutrient concentrations, it was suspected that this dramatic increase in the plankton might reflect the relief of nutrient-limited growth through anthropogenic inputs. However, when these CPR averages for the North Sea were resolved into component rectangles of 1° latitude by 2° longitude to identify the site of maximum change, it was immediately apparent that such a cause was unlikely.

For each such rectangle, Figure 4a illustrates the average increase in *C. fusus, C. furca,* and *C. tripos* in the period 1982–1988, cf. the long-term mean 1958–1981. Areas of decrease are hatched. It shows that the locus of maximum increase (17-fold compared with the long-term mean) lay east of the Shetlands, with a tongue of lesser increase running down the centre of the northern North Sea, west of the Norwegian Trench. Figures 4b and c confirm that essentially the same geographic pattern of increase was followed by other components of the summer plankton (*C. macroceros, C. longipes, Pseudocalanus* and *Limacina* in Fig. 4b; Chaetognatha in Fig. 4c). Plainly, although it is possible to identify *some* components of the plankton which exhibit *some* increase further south, we are dealing here with a phenomenon of the northern North Sea, remote from the zone of maximum anthropogenic nutrient loadings along the continental coast of the German Bight.

Figure 5, adapted from Robinson (1961), provides an immediate clue as to the true cause of the *Ceratium* peak in showing that these species are normally located in summer in that part of the northeastern North Sea that is most affected by the freshwater outflow from the Baltic. The cause is not hard to seek. Ceratia are large with a high rate of respiration to photosynthesis (26% cf. 7% for diatoms; Harris, 1978; Cushing, 1989) and a slow daily division rate (0.2–0.4 d^{-1} cf. 1–2 d^{-1} for diatoms). They therefore recover much more slowly than other components of the plankton if they are disturbed from their preferred near-surface depth zone by (for example) storm-mixing. Though they could regain their preferred depth by swimming, their swim-rate is slow (~0.5 m h^{-1}). Thus "the depth of the pycnocline is probably critical in maintaining a suitable light and nutrient environment for growth" (Holligan, 1987, p. 14).

For these sorts of reasons they "demand permanent stratification" (Cushing, 1989, p. 4) and bloom at a time and place where permanent stratification is most dependably found. Within the Baltic outflow zone in summer the effects of a seasonal thermocline and a permanent halocline reliably combine to establish a strong near-surface density gradient as a barrier to vertical mixing.

While turbulence on many scales appears detrimental to dinoflagellate division rate (see Pollingher and Zemel, 1981), dinoflagellates have positive advantages over other components of the plankton when located in conditions of high water column stability. Being large, and of a slow division rate, the Ceratia leak less exudate than smaller cells (Raven, 1986; Fogg, 1986); since Ceratia may develop a biomass equivalent to 1000 mg Cm^{-3} (Radach *et al.*, 1990), it is important that their photosynthetic apparatus has means of coping with the self-shading that a successful bloom in stratified conditions may engender (Harding *et al.*, 1983; Holligan, 1987); and their motility permits marine Ceratia to vary their diel migration through the thermocline according to their requirements for nutrients and light, with a progressively earlier downward descent during daytime, and a longer and more thorough penetration of the thermocline as nutrient depletion sets in (Heaney and Eppley, 1981). Given stratification then, Ceratia are better equipped than most to prosper. The question, then, is whether reinforced stratification was the cause of the enhanced *Ceratium* bloom in the north-central North Sea during the summers of 1987–1988.

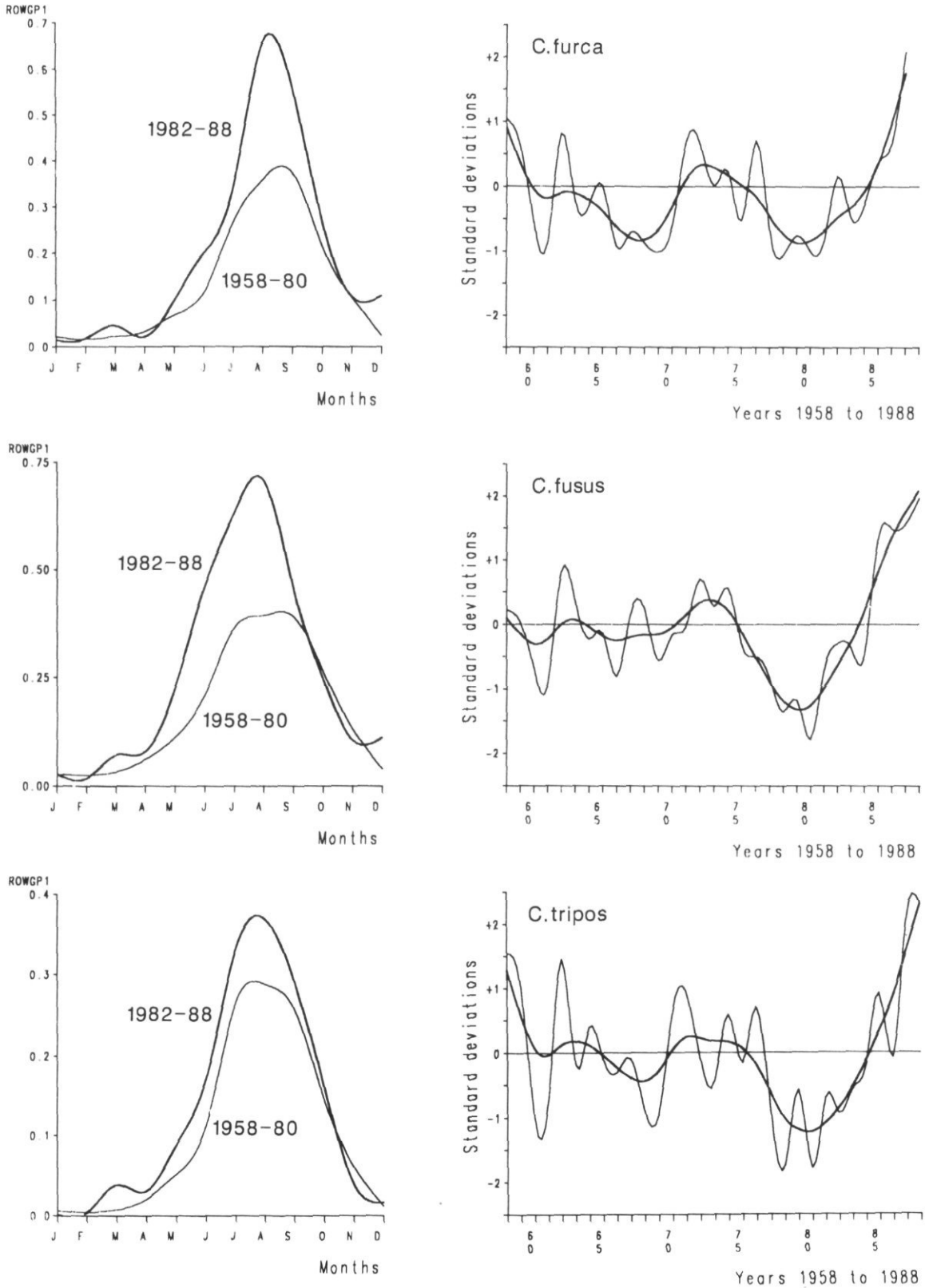


Figure 1. North Sea averages of the annual fluctuation in abundance (standardized to zero mean and unit variance) for the three *Ceratium* species, *C. furca*, *C. fusus*, and *C. tripos*, 1958–1988 (CPR data).

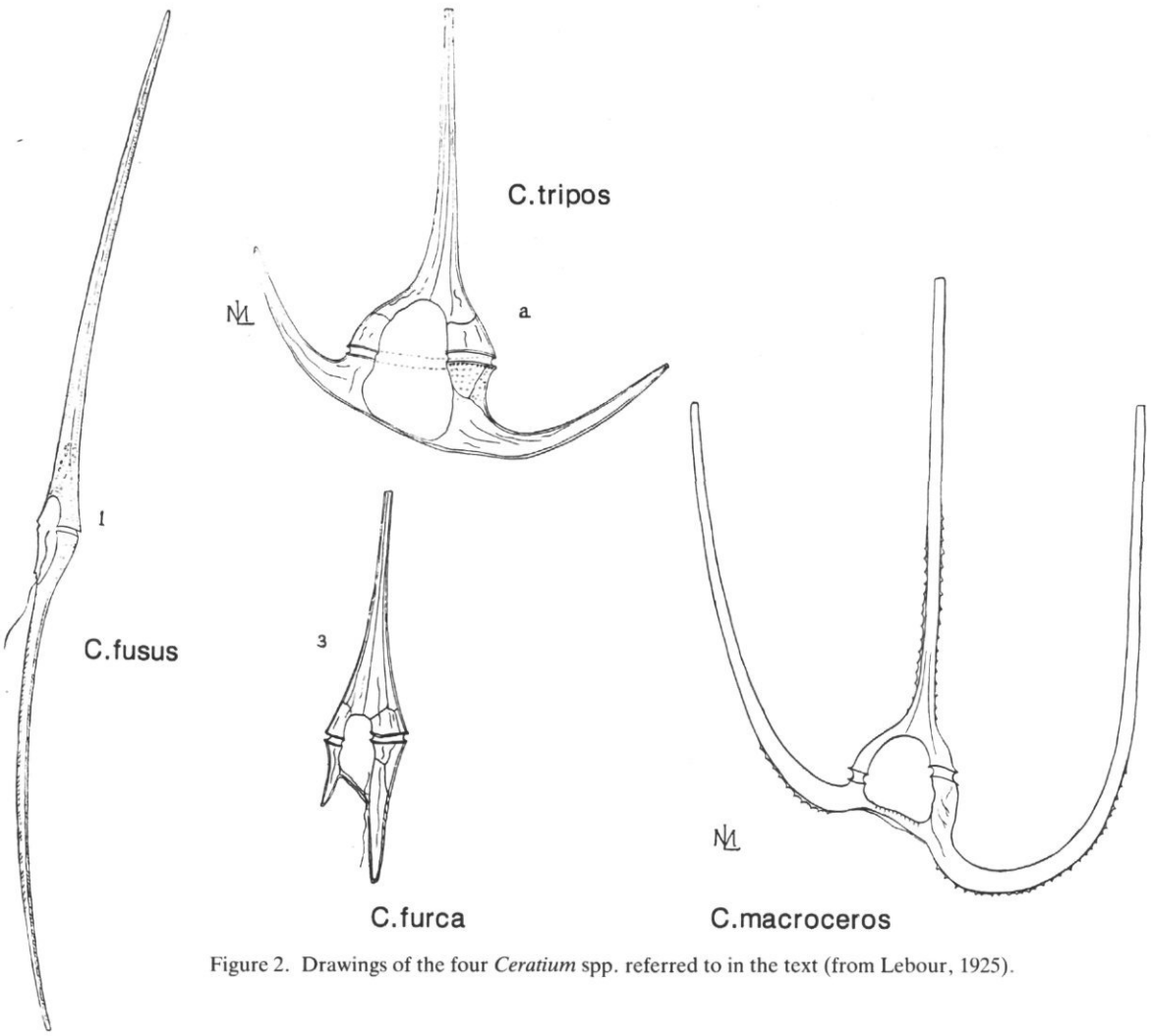


Figure 2. Drawings of the four *Ceratium* spp. referred to in the text (from Lebour, 1925).

The physical environment 1987–1988

Although thermal stratification is present right across the northern North Sea in summer, Ljoen's (1980) Atlas confirms that the strongest density gradients in the near-surface zone are determined by salinity stratification and are therefore found in the Skagerrak and along the west coast of Norway in the narrow zone occupied by the Norwegian Coastal Current. When we compare the long-term average summer density distribution on the Utsira–Orkney Section (1967–1976) with that encountered by RV "G. O. Sars" on 12–14 July 1987, it is plain that the summer 1987 conditions are grossly anomalous (Fig. 6). Near-surface isopycnals have spread west from their normal location close to the Norwegian Coast, with $\sigma t = 26$ almost reaching the prime meridian, and there has been a corresponding reinforcement of the pycnocline at 10–30 m depth.

To confirm the regional extent of this anomaly, the results from all of the Norwegian standard hydrographic sections were used to map the near-surface (0–30 m) mean salinity distribution in summer 1987, and this distribution is compared with the long-term 1967–1976 mean (Ljoen, 1980) in Figure 7.

The comparison shows an unprecedented westward bulge of fresh coastal water across the northern North Sea from its normal constrained location around the coasts of southwest Norway, with the 35 isohaline now lying close to Shetland instead of in the central North Sea and salinities <34 extending to west of 1°E. A repeat analysis for summer 1988 (not shown) shows an almost equally extreme westward shift, although with the main dislocation lying further south, west of the Skagerrak.

When we now overlay the pattern of increased *Ceratium* abundance (from Fig. 4a) onto a chart representing the freshening of the near-surface layer between 1967–1976 and 1987 (Fig. 8), the hypothetical relationship between the reinforcement of the pycnocline and the amplitude of the *Ceratium* bloom appears confirmed.

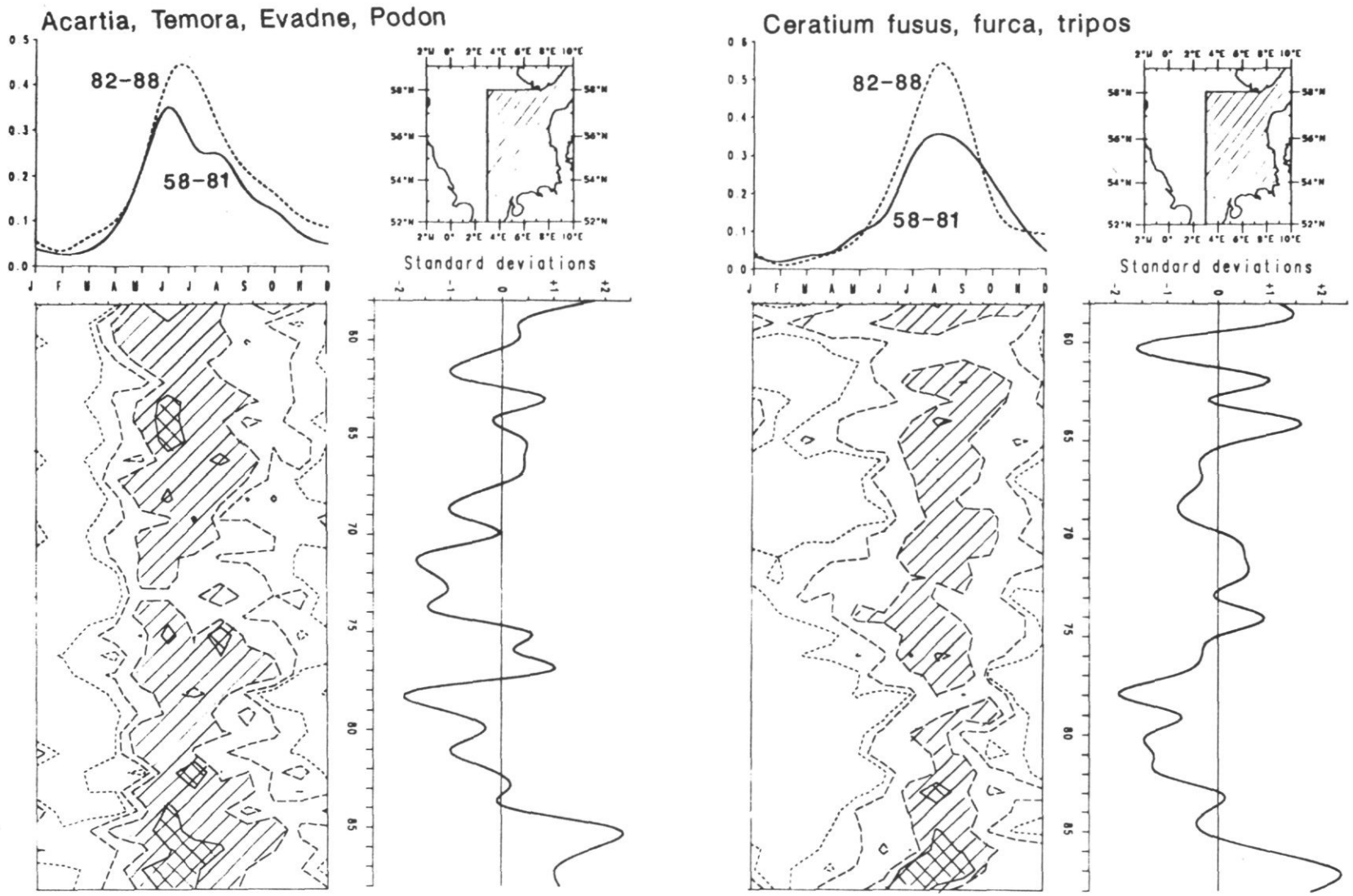


Figure 3. Annual fluctuations in abundance of the Ceratia *C. fusus*, *C. furca*, and *C. tripos*, the neritic copepods *Acartia* and *Temora* and the Cladocerans *Evadne* and *Podon* in the eastern North Sea, 1958–1988 (CPR data).

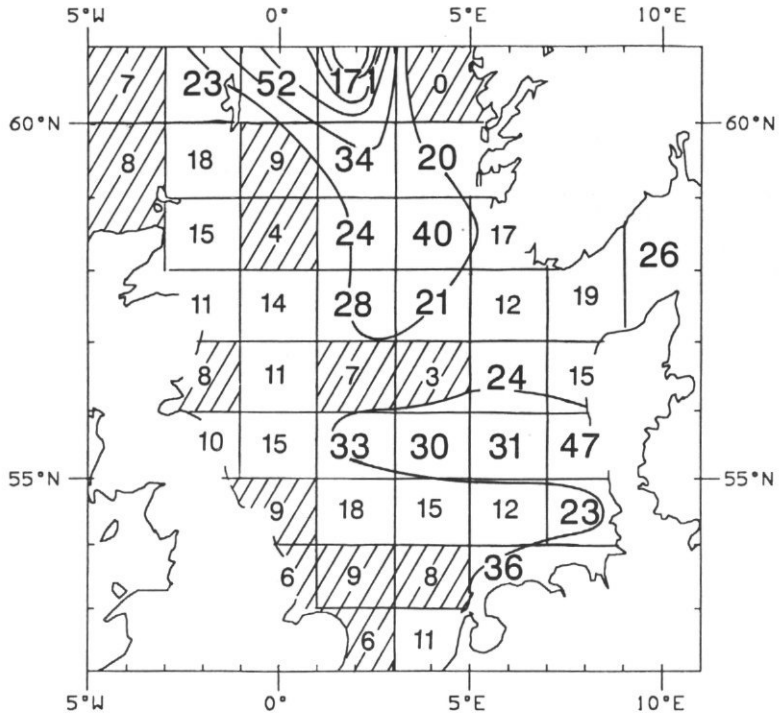


Figure 4a. Distribution of the average increase (×10) of *C. fusus*, *C. furca*, and *C. tripos* in the period 1982–1988, compared with their long-term mean abundance, 1958–1981 (i.e: A value of 30 signifies a three-fold increase; areas of decrease are hatched).

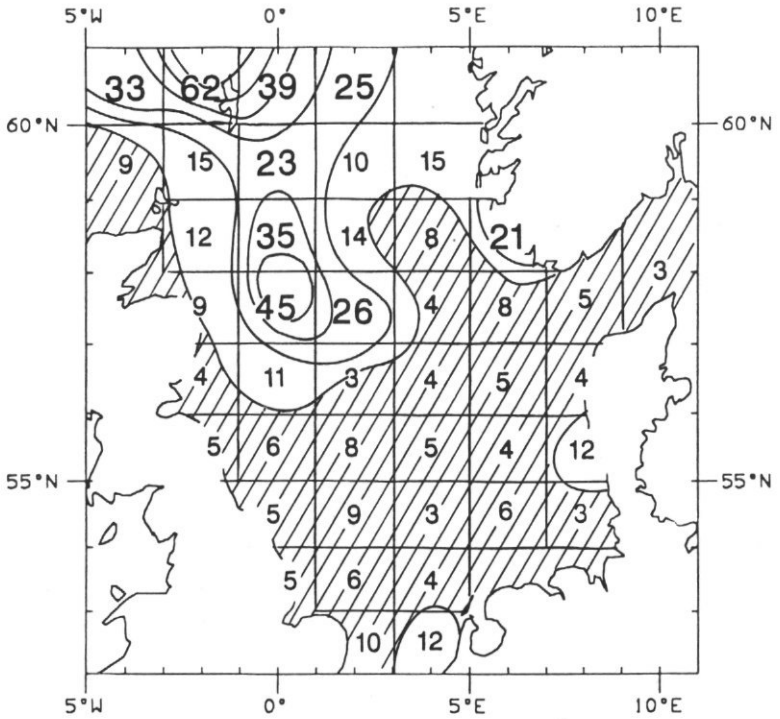


Figure 4b. Distribution of the average increase (×10) of *C. macroceros, C. longipes, Pseudocalanus*, and *Limacina* in the period 1982–1988, compared with their long-term mean abundance, 1948–1981. (A value of 45 signifies a 4.5-fold increase; areas of decrease are hatched.)

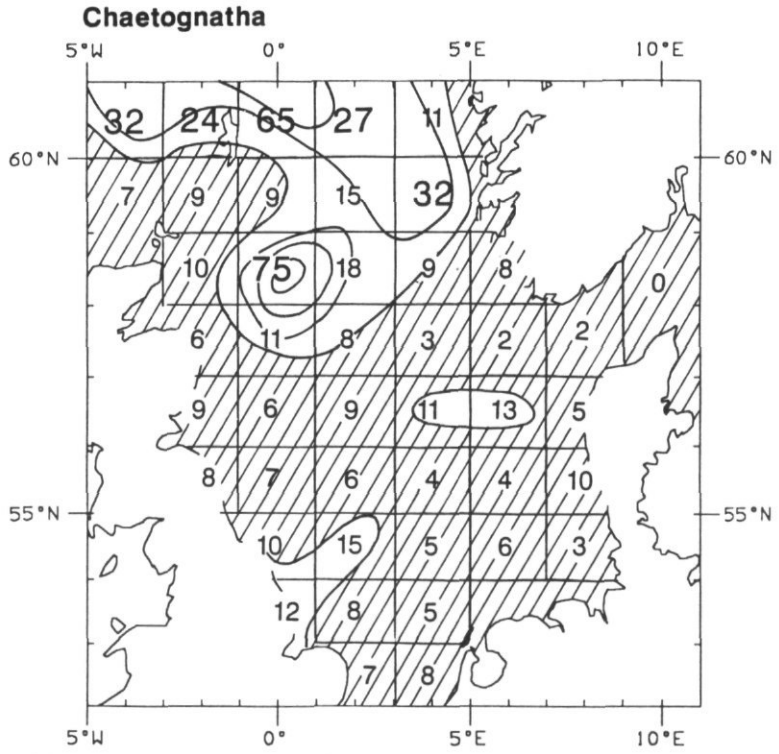


Figure 4c. Distribution of the average increase (×10) of Chaetognatha in the period 1982–1988, compared with their long-term mean abundance, 1948–1981. (A value of 75 signifies a 7.5-fold increase; areas of decrease are hatched.)

However clear the relationship may be in this instance, the question remains as to whether we are dealing with a one-off "event" driven by a singularity, or with an exaggeration of the same set of factors which normally determine *Ceratium* abundance. Figure 9 appears to suggest the former in indicating that over the period 1968–1988, there has been no *general* relationship between *Ceratium* abundance and the areal spread of surface (0–30 m) water of salinity <34.0 in the northern North Sea, other than in the years under discussion. Thus multiple regression analysis would identify a range of environmental parameters as contributing to the normal year-to-year variability of *Ceratium* in this area, including a range of antecedent wind variables, cloud cover, etc., but does not select for surface "fresh"-water extent. The explanation appears to be that the origin as well as the extent of the coastal water is of importance, and *that* is dependent on wind direction.

In modelling the shelf-seas circulations that arise due to steady wind stress for varying wind directions, Pingree and Griffiths (1980) show that the result of a uniform southeasterly wind stress is a strong Skagerrak outflow and a spreading of coastal water across the north-central North Sea, much as we observe here. And if we compare the mean summer atmospheric circulation pattern of the late 1980s (when the *Ceratium* bloomed) with that of the early 1980s, we find that the years of the *Ceratium* bloom were indeed dominated by an intensified southeasterly airflow sweeping the northern North Sea (Fig. 10). The detachment and westward spreading of fresh surface water from the Norwegian coast is, on the other hand, a function of northerly wind stress, especially in May and June (Saetre *et al.*, 1988), and the multiple regression analysis by Svendsen and Magnusson (1991) leaves little doubt that this is the mechanism which has *normally* determined the summer distribution of coastal water in the northern North Sea over the period 1968–1990.

The extreme extent of coastal water across the northern North Sea in summer 1987 and the widespread stable surface layer that resulted are thus mainly attributed to enhanced Skagerrak outflow in that summer and in the preceding March when easterly winds were particularly strong. To a lesser extent their effect would have been compounded by a westward spread of Norwegian coastal water in the intervening months of May and June when winds were weak but rather persistently from the North.

Discussion and conclusions

Thus our hypothesis is a simple one, attributable to a change in local climate. *Ceratium* species reached unprecedented summer abundance in the north-central North Sea during the late 1980s because the highly stratified

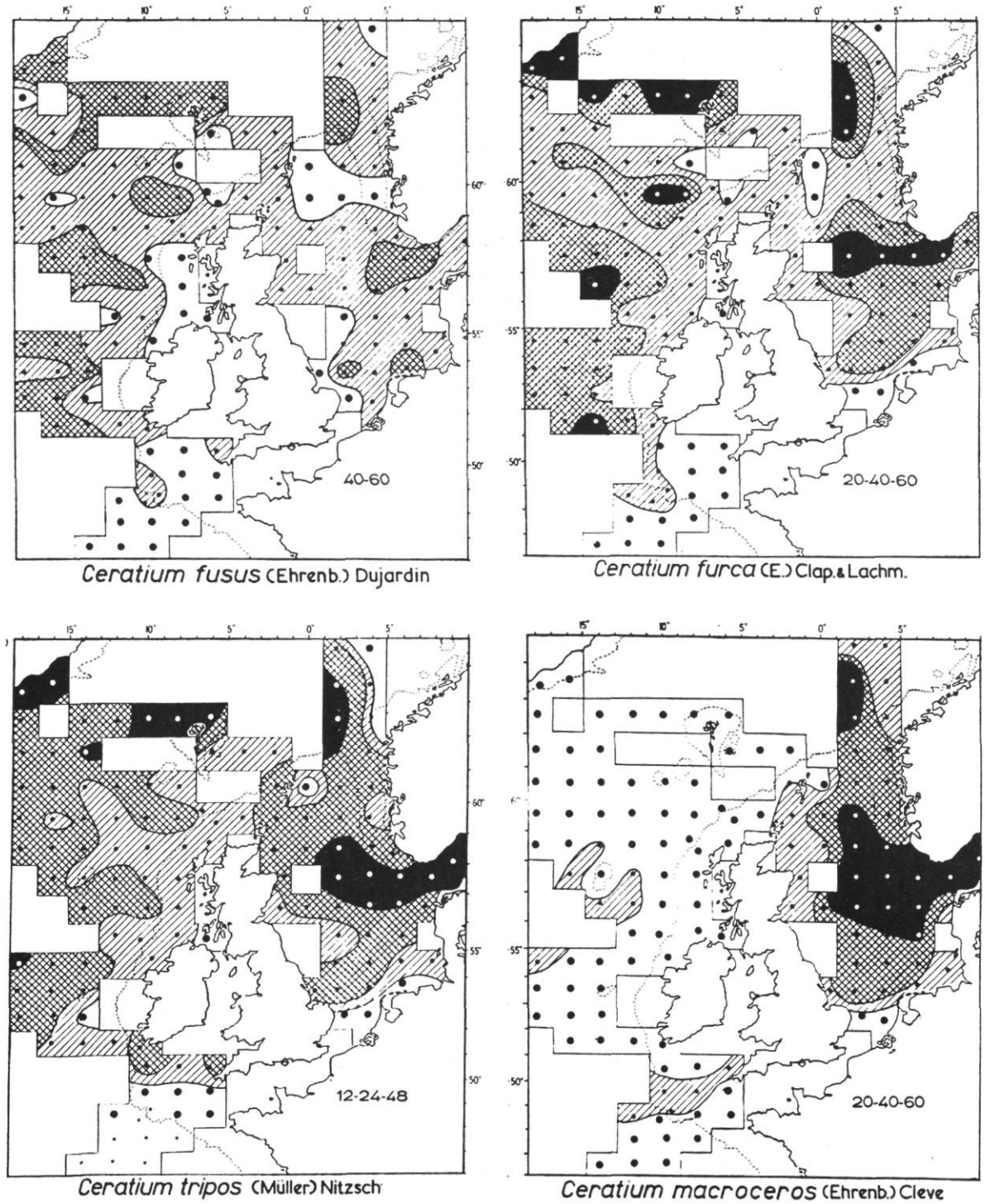


Figure 5. Mean distributions of *C. fusus*, *C. furca*, *C. tripos*, and *C. macroceros* in the northeast Atlantic and European shelf, 1948–1956 (from Robinson, 1961).

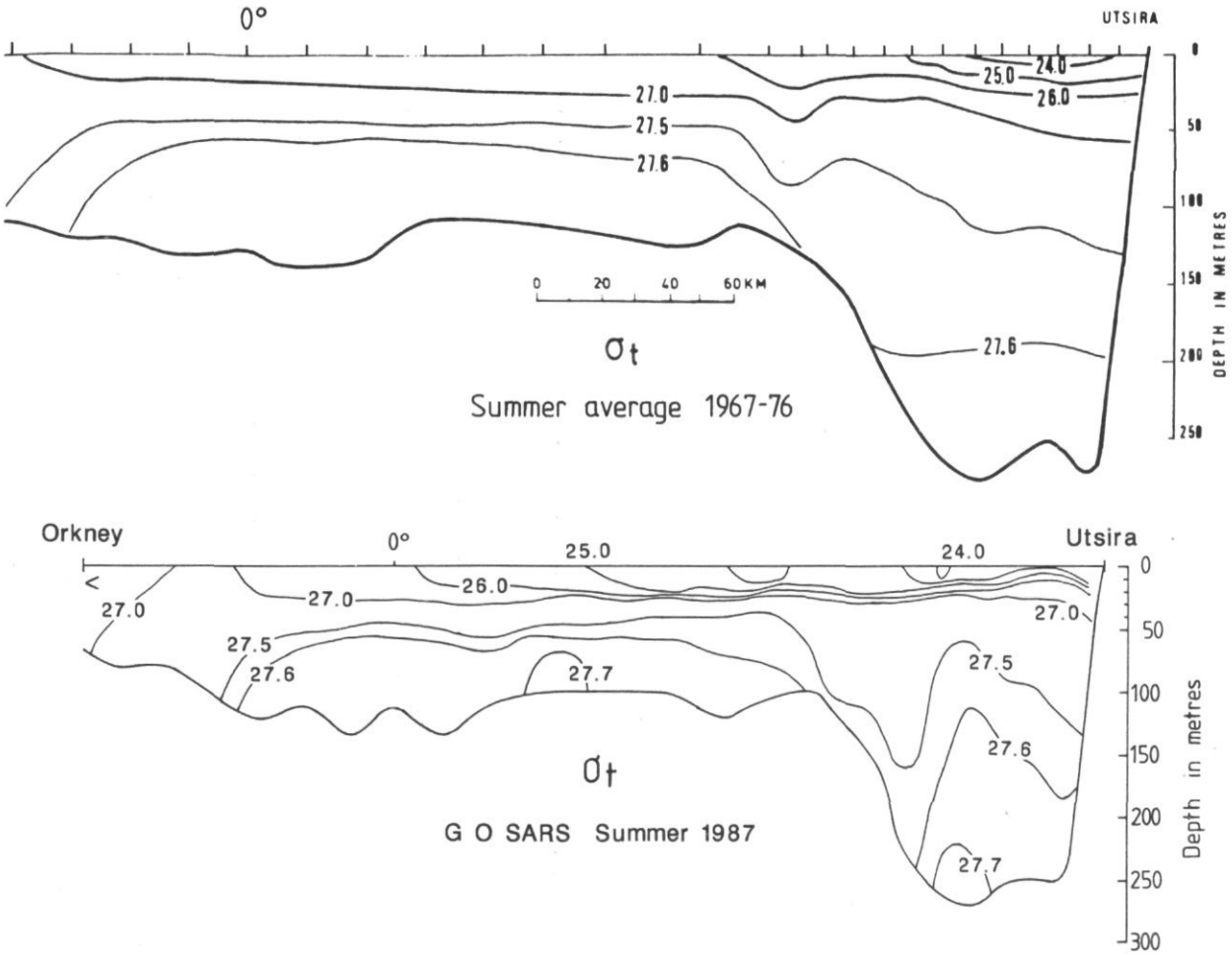
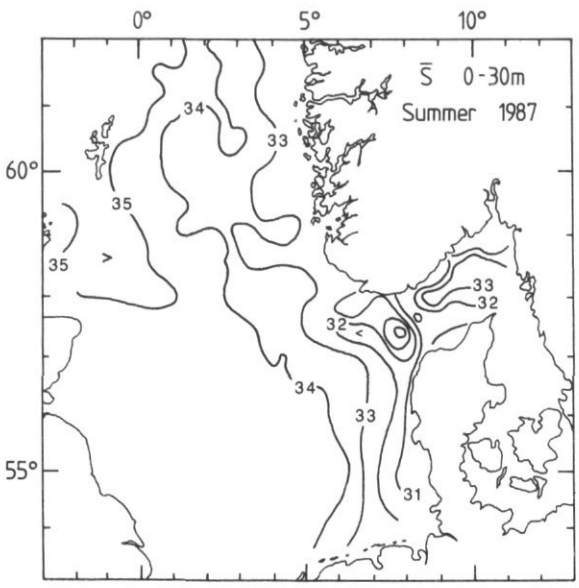


Figure 6. Distribution of isopycnals on the Utsira–Orkney standard hydrographic section on 12–14 July 1987, compared with the long-term average distribution for summers 1967–1976 (from Ljoen, 1980).



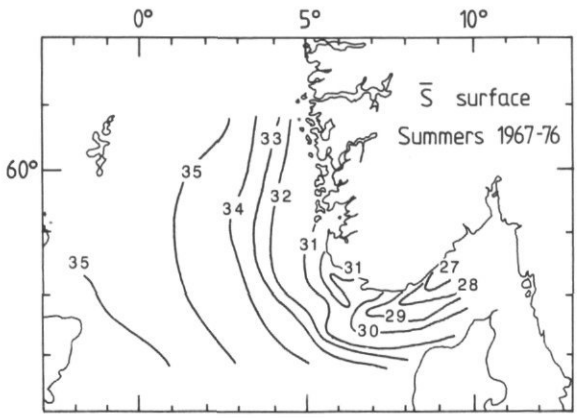


Figure 7. Horizontal distribution of mean surface (0–30 m) salinity in the North Sea during summer 1987, compared with the long-term summer mean, 1967–1976 (from Ljoen, 1980).

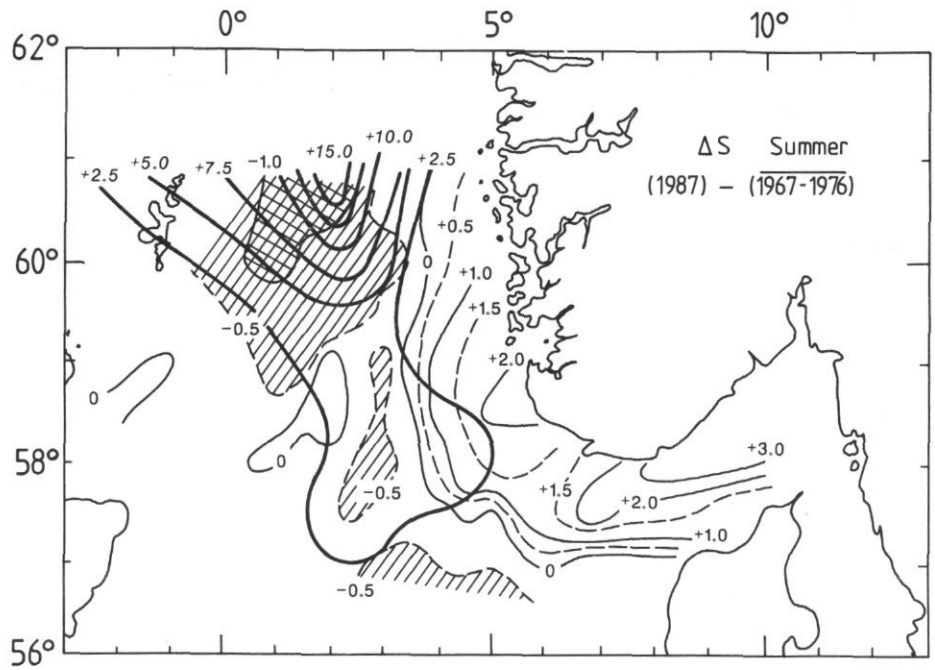


Figure 8. Distribution of freshening in the surface (0–30 m) layer of the northern North Sea in summer 1967, compared with the long-term (1967–1976) mean, with contours of increased *Ceratium* abundance superimposed. ("+10" signifies a 10-fold increase in abundance in 1982–1988, cf. the long-term mean abundance, 1958–1981).

conditions that they not only like but *require* were brought westwards across that area under the stress of an intense easterly airstream. A similar (but more limited) westward extension of fresh surface water from the Jutland Coast may have supported a similar but less dramatic increase of *Ceratium* spp. in the German Bight

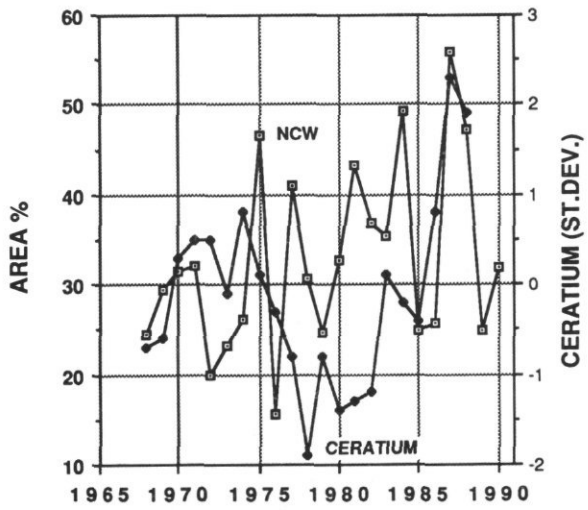


Figure 9. Time series of North Sea *Ceratium* abundance, compared with the changing areal distribution of Norwegian Coastal Water, 1968–1990. ("NCW" is defined as surface (0–30 m) water with a mean salinity <34.0, and its distribution is measured by its percentage coverage in the northern North Sea between 57°N and 60°50'N (northern tip of Shetland).)

(see Fig. 4a). The role of anthropogenic eutrophication is discounted in this particular case. Even though Skagerrak surface water may be enriched relative to that of the northern North Sea, Kiørboe *et al.* (1990) show that these "excess" nutrients usually become rapidly and locally depleted, but we lack reliable nutrient time series from the northern North Sea to check this point in this instance.

The parallel increase which the CPR observed in the abundance of the Cladocerans *Evadne* and *Podon* almost certainly has the simplest of explanations. In Danish coastal waters where *Ceratium* dominate the summer phytoplankton, Nielsen (1990) provides clear circumstantial evidence that Cladocerans are one of the few groups able to graze *Ceratium*, removing 30–70% of the *Ceratium* production per day, with Cladoceran vertical distribution correlating with of *Ceratium* and with the Cladoceran biomass increasing until it "totally dominated the mesozooplankton" while the large copepod biomass remained constant. As Nielsen concludes (p. 29) "The most important grazers, the heterotrophic dinoflagellates and the Cladocerans have not received much attention so far, but these groups alone possess the potential of controlling the development of *Ceratium* blooms, so future investigations have to include them if the dynamics of the *Ceratium* blooms have to be fully understood". In the present context, Cladocerans showed a parallel increase because they eat *Ceratium*.

Whether or not the above explanation is valid, the reversion to normal *Ceratium* abundance again in the



Figure 10. Mean change in surface pressure between summers (1980–1984) and summers (1985–1988).

northern North Sea during the summer of 1989 is the clearest indication that we are dealing here with the local effects of short-term climatic "noise", not with the sustained trends of change that we might expect from slow shifts of global climate or regional contamination.

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Continuous plankton records: the North Sea in the 1980s

The CPR Survey Team

CPR Survey Team, the. 1992. Continuous plankton records: the North Sea in the 1980s. – ICES mar. Sci. Symp., 195: 243–248.

The results of the Continuous Plankton Recorder Survey are used to describe changes in the plankton of the North Sea during the 1980s in relation to the three previous decades. Coherent patterns of change in abundance, in the form of a decline to the early 1980s and a subsequent recovery, and in the relationships between species can be related to a changing climatic syndrome associated with global warming. The main pattern of change can be observed across the four trophic levels, phytoplankton, zooplankton, fish, and sea birds; and in a variety of physical phenomena in the North Atlantic and beyond.

The CPR Survey Team: The Sir Alister Hardy Foundation for Ocean Science, c/o Plymouth Marine Laboratory, Prospect Place, The Hoe, Plymouth PL1 3DH, UK.

Dickson *et al.* (this volume) describe some recent changes in the abundance of several species of *Ceratium* in the North Sea. It takes the form of a description of an event restricted to a few species in the plankton, delimited in time to the summers of a couple of years and in space to the northern North Sea. The paper also includes a very plausible explanation of the event in terms of the biology of the organisms involved and in relation to hydrography and climate.

This paper looks at the same event, using the same data, from the Continuous Plankton Recorder (CPR) survey, but from the point of view of seeing what happens when it is set in the context of a decadal time scale. As the starting point, Figure 1 moves from a couple of summers in the northern North Sea to the whole world for the whole of this century: it shows the changes in global air temperature (Pearce, 1991) and indicates that there has been a fairly steady warming through most of the 20th century.

The data from the CPR survey only covers the period from 1948 and Figure 2a shows only this part of the time series. The temperatures stayed more or less constant, with no apparent trend, for the first 25 years or so. It is only since the mid-1970s that the warming trend apparent during the first few decades of the century has resumed.

Taking a step closer towards the plankton, Figure 2b shows the sea-surface temperature for the North Atlantic Ocean (Folland and Parker, 1990). Compared with the air temperatures, the sea temperatures were actually

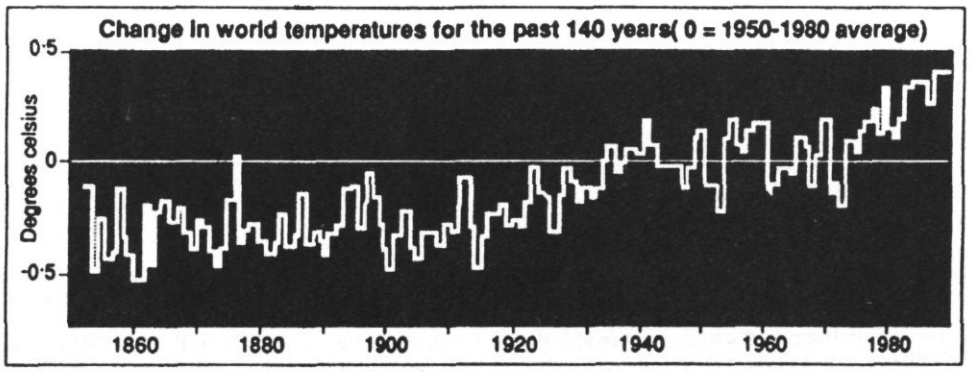

Figure 1. Graph of global mean surface air temperatures from 1850 to the present. Reproduced from Pearce, 1991.

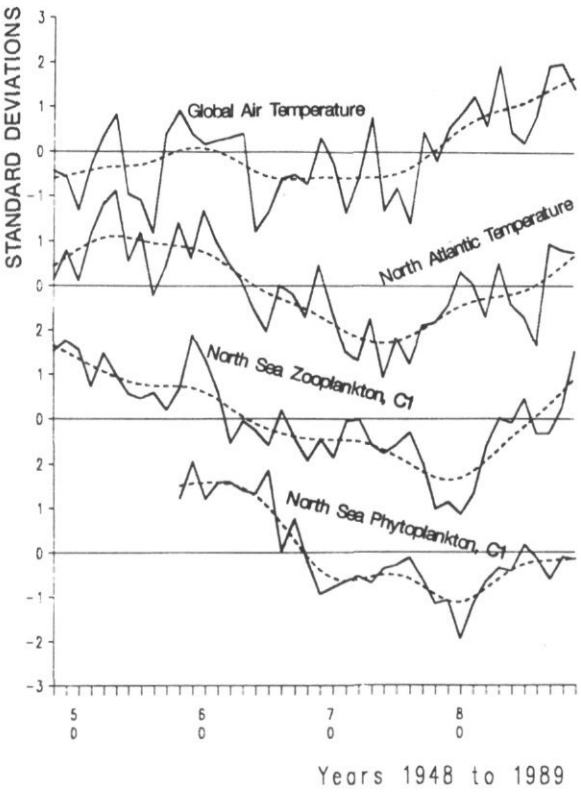

Figure 2. Standardized (zero mean and unit variance) annual time series and fitted smoothed (cubic spline) curves of: (a) Global air temperature (Kerr, 1990), (b) North Atlantic sea-surface temperature (Folland and Parker, 1990), (c) and (d) first principal components of the fluctuations in the abundance of zooplankton and phytoplankton in the North Sea (CPR survey).

falling during the 1950s and 1960s. The global air temperature and the Atlantic surface temperatures share a common warming trend only since the mid-1970s.

Figures 2c and d add the plankton to the picture. These two graphs are representations of changes in the abundance of zooplankton and phytoplankton in the North Sea. The phytoplankton data start in 1958 because the method of counting was changed then and the earlier data are not comparable. These two variables are Principal Components: they are the best possible single representations of the annual fluctuations of 19 species of zooplankton and 21 species of phytoplankton respectively, based on the criterion of maximum variance. Both show a decline in abundance followed by a recovery. The pattern is similar to the Atlantic temperatures except that while the Atlantic started to warm up in the mid-1970s, the plankton did not begin to recover until about 1980.

There is reason to believe that the common pattern shown here is more than coincidence. Although the details of the processes and mechanisms involved are far from clear, the long-term pattern of change in the abundance of the plankton is firmly tied to long-term changes in climate. There are currently two stories relating to changes in the intensity of either northerly winds or westerly winds. The North Wind story (Dickson *et al.*, 1988) is the more highly developed from the point of view of interpretation in that it can be related to the timing and intensity of the spring bloom of phytoplankton. The West Wind story (Colebrook, 1985) is statistically more firmly based but all that is known about the processes is that they appear to happen in winter and, unfortunately, not much is known about what happens to the plankton in the winter-time. It is probable that in both cases the same basic pattern of climatic change is involved. The uncertainty relates only to the processes linking the climate to the plankton. It is not impossible that both stories are right. What is fairly certain, however, is that the changes in the plankton are not a direct response to changes in temperature and that the pattern of change in the North Atlantic (Fig. 2b) should be considered, from the point of view of the plankton, simply as an index of climatic change.

This being so, it can be asked whether there have been any changes in the plankton that date from the mid-1970s and might have acted as a precursor to the increase in abundance from the early 1980s. Figure 3 repeats the first components for the plankton (a and c), which represent 32% and 36% of total variability respectively. The figure also includes the second components (b and d). These can be considered as representing residual patterns of change in abundance and account for 16% and 18% respectively of total variability. For comparison with the second components Figure 3 also shows plots of the annual means of abundance of the three species of *Ceratium* (e, f, and g) included in Figure 1 of Dickson *et al.* (this volume) together with plots for *Evadne* and *Podon* which feed on the *Ceratium* species. It is clear that the second component for phytoplankton is reflecting the *Ceratium* story and, at least partly through a trophic link, this is represented in the zooplankton as well.

It is inherent in Principal Components Analysis that pairs of components are orthogonal. They are, by definition, uncorrelated. But, looking at the pair of zooplankton components it is clear that from the mid-1970s they are correlated; also for the phytoplankton, although less pronounced. For this to happen there has to have been a change in the form of the relationships between the species over this period. This can be considered as representing a change in the community structure, although it is by no means certain what community means in the context of the plankton.

For the zooplankton, it is possible to split the data at 1975 and to do separate analyses on the two sectors. The scatter plots in Figure 4 are vector diagrams representing, as far as this is possible in two dimensions, a spatial analogue of the relationships between the species

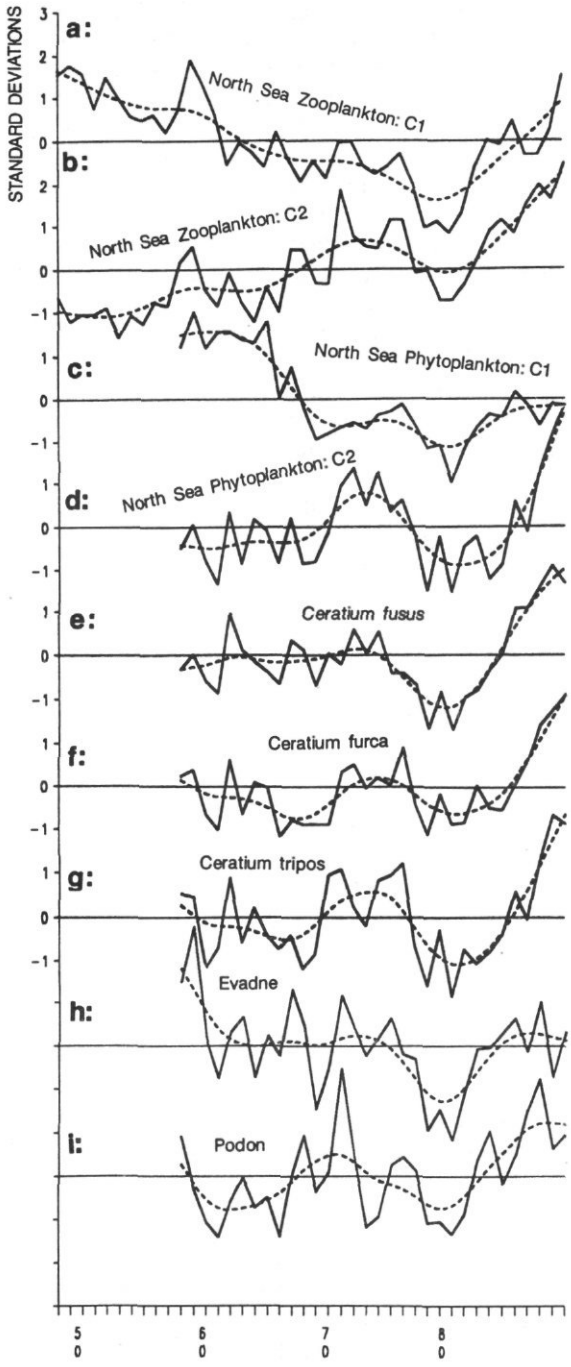

Figure 3. Standardized annual time series and fitted smoothed curves of annual fluctuations of abundance in the North Sea of: zooplankton (a and b) and phytoplankton (c and d), *Ceratium fusus, furca* and *tripos* (e, f and g) and *Evadne* spp. and *Podon* spp. (h and i).

with respect to their annual fluctuations in abundance. Species close together are similar, those further apart are different. For technical reasons the number of species included in each analysis had to be reduced from 19 to 10. It is obvious that the pattern of species in the two plots are different. In the period after 1975 there is a larger coherent group of species at the positive end of the first vector axis as compared with 1948–1974. The position of this group indicates that they all contribute positively to the first component. It is worth noting that two of the four zooplankton species mentioned by Dickson *et al.* (this volume) as being involved in the *Ceratium* story are included; they are *Acartia* (Ac) and *Podon* (Pd) and both have moved into the large coherent group.

Of particular interest are the two species *Acartia clausi* (Ac) and *Pseudocalanus elongatus* (Ps). Prior to

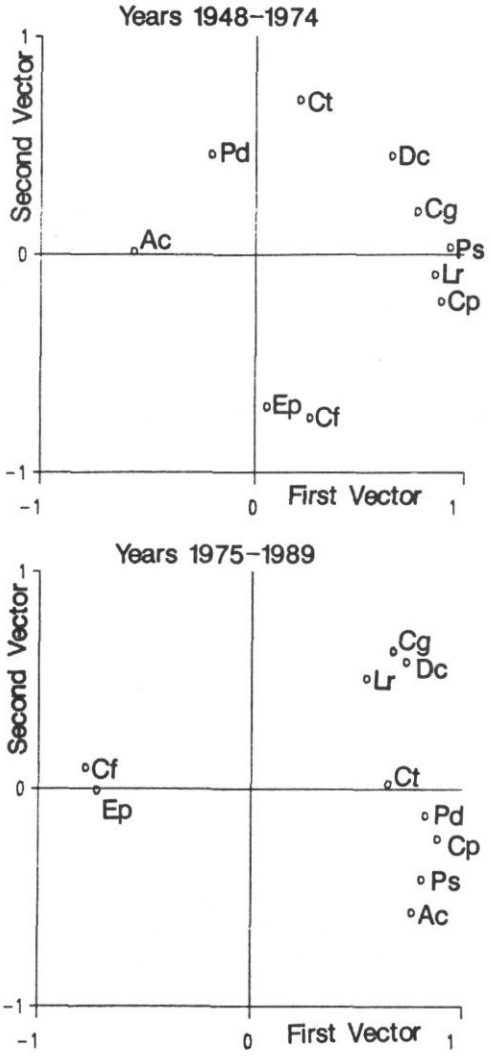

Figure 4. Plots of the first two vectors of principal components analyses of the annual fluctuations in abundance of ten taxa of zooplankton in the North Sea for the periods 1948–1974 and 1975–1989 respectively. Key to the taxa: Ac – *Acartia clausi*, Pd – *Podon* spp., Ct – *Centropages typicus*, Dc – *Decapod larvae*, Cg – *Chaetognaths*, Ps – *Pseudocalanus elongatus*, Lr – *Limacina retroversa*, Cp – *Total Copepods*, Cf – *Calanus finmarchicus*, Ep – *Euphausiacea*.

1975 these two species appear at more or less opposite ends of the first vector axis, while after 1975 they appear close together, positively correlated. It so happens that something is known about the life-histories of these two species. *Pseudocalanus elongatus* is relatively slow growing with a relatively low fecundity. Its abundance and distribution are determined to a large extent by the range and size of its overwintering stocks, because it cannot grow and reproduce itself fast enough for the summer populations to achieve independence from the winter stocks. *Acartia clausi*, on the other hand, can grow and reproduce faster, not all that much faster, but sufficiently for the summer populations to achieve a measure of independence from the overwintering stocks.

These two species are reasonably abundant and sampled well enough by the CPR to permit resolution of year-to-year changes on a monthly basis. That is, to look at the time series for all the Januaries and all the Februaries and so on. Figure 5 contains correlation matrices, based on bi-monthly data to simplify the presentation, for: within *Pseudocalanus*, within *Acartia*, and between the two species, for each of the periods 1948 to 1974 and 1975 to 1989.

Looking first at *Pseudocalanus*, the correlations are slightly higher between successive months, but not much higher than elsewhere in either matrix. There is no clear indication of systematic seasonal differentiation of the pattern of year-to-year fluctuations in abundance. For *Acartia*, and for the period 1948 to 1974, the story is different. The correlations are all low, indicating very little consistent pattern in the year-to-year fluctuations between successive pairs of months, whereas, for 1975 to 1989, there is clearly greater coherence through the year. Although not as high as for *Pseudocalanus*, the correlations are noticeably higher than for the early period.

Comparing *Pseudocalanus* with *Acartia*: for 1948 to 1974 what positive correlations there are occur in March/April, while in summer they are mostly small and negative. From 1975 to 1989 the position is very different, nearly all the correlations are positive and, in particular, the correlations between corresponding month pairs are high. The implication is fairly obvious, that prior to 1975 the processes resulting in the pattern of decline in abundance operated chiefly in winter and early spring, and manifest in both species only in March/April, while in the later period, from 1975 to 1989, it would appear that the processes have been operating throughout the year, or at least well into the summer, reducing seasonal differentiation in the pattern of year-to-year changes.

Although it can only be demonstrated for these two species, it seems reasonable to assume that this phenomenon is responsible for the generally greater coherence between the species in the period from the mid-1970s.

Because of the shorter time series, and because the winter stocks of phytoplankton are very low, it is not possible to repeat this kind of analysis for the phytoplankton. Nevertheless, the similarity between the component pairs of zooplankton and phytoplankton suggests that the same kind of thing has happened there. That is, in the period from the mid-1970s there has been a reduction in seasonal differentiation in the pattern of year-to-year changes resulting in a greater coherence between the species.

Going back to the original intention of seeing how the plankton/climate event described by Dickson *et al.* (this volume) fits into changes taking place on a longer time scale, it would appear that the event can indeed be seen as part of a longer story; that what can be described and explained as a response to a relatively transient anomaly in the wind-field can also be seen as part of a wider response to a changing climatic syndrome.



Figure 5. Correlation matrices (values as correlation coefficients × 10) for the annual fluctuations in the abundance of *Pseudocalanus elongatus* and *Acartia clausi* in sets of pairs of months for the North Sea for the periods 1948–1974 and 1975–1989 respectively.

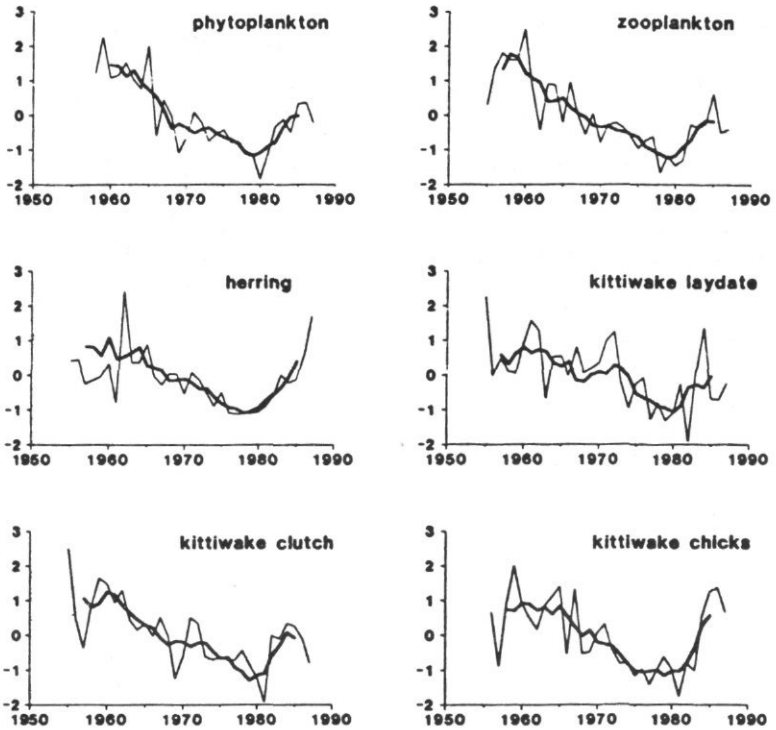

Figure 6. Standardized time series and five-year running means for 1955–1987 of: abundances of phytoplankton and zooplankton for the northeast North Sea; numbers of two-year-old herring for the northern North Sea; and, for a colony of kittiwakes on the northeast coast of England, the laying date, clutch size and chicks fledged. Data from Aebischer, Coulson, and Colebrook, 1990.

So far, only plankton/climate relationships in the North Sea have been considered. Even in the North Sea the pattern of change shown in Figure 2 is not confined to the plankton; as is shown in Figure 6, it occurs across four trophic levels, phytoplankton, zooplankton, herring stocks, and seabirds. Moving outside the North Sea, North Atlantic temperatures have already been considered and are repeated in Figure 7b. To this can be added changes in the latitude of the north wall of the Gulf Stream off Cape Hatteras (Fig. 7a, Taylor, pers. comm.) and upwelling off the coast of Portugal (Fig. 7c, Bakun, 1990). These are compared with the first component for zooplankton (Fig. 7d). Staying in the same hemisphere but moving into the Pacific Ocean, the same pattern can be seen in upwelling off the coast of California (Fig. 7e, Bakun, 1990) and also in estimates of stocks of planktonic Diatoms and Foraminifera based on analyses of bottom sediments in the Santa Barbara Basin by Lange *et al.* (1990), (Fig. 7f and g). Clearly, these are only fragments, but hopefully sufficient to indicate the probability of the existence of a coherent hemispheric, if not global, story.

The value of events of the kind described by Dickson *et al.* (this volume) is that they provide opportunities for explanation in terms of relationships between climate and hydrography and between hydrography and the biology of the organisms involved. Whereas, in terms of the larger story we are still at the stage of trying to identify and simply describe the factors involved.

Finally, coincident with the change in timing of the *Ceratium* species (Fig. 8a) there has been a marked increase in the abundance of two copepod species normally considered to be indicators of the incursion of Atlantic water into the North Sea, *Metridia lucens* (Fig. 8b) from the north, and *Corycaeus anglicus* (Fig. 8c) through the English Channel. Also in 1989, the presence was recorded of the doliolid *Doliolum nationalis* just north of the Dogger Bank and in the Southern Bight (Lindley *et al.*, 1990). This species has not been recorded anywhere in the North Sea since 1912. The interpretation in terms of Atlantic inflow is supported by observations of exceptionally high salinity water in the northern North Sea in the winter of 1990 (Heath *et al.*, 1991). There is not necessarily any conflict between a high Atlantic inflow and the spread of low salinity water from the Skagerrak. The Atlantic inflow appears to be occurring in autumn and winter, while the high abundance of *Ceratium* is a summer phenomenon.

Going back to Figure 1, it is implied that what is happening in the North Sea is part of a global pattern of climate change. It is also implied that such changes are taking place in the context of a pattern of global warm-

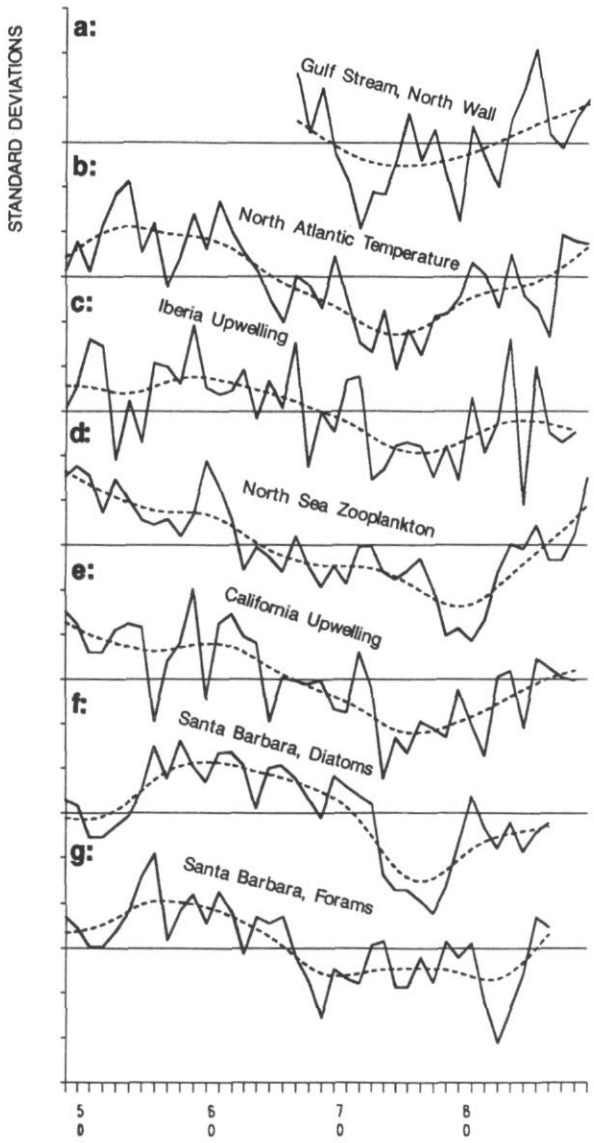

Figure 7. Graphs of standardized time series and smoothed curves of: a – an index of the latitude of the north wall of the Gulf Stream (data from A. Taylor, pers. comm.), b – North Atlantic temperature (repeated from Fig. 2), c – an index of coastal upwelling off Iberia (Bakun, 1990), d – North Sea zooplankton (repeated from Fig. 2), e – an index of coastal upwelling off California (Bakun, 1990) estimates of the abundance of planktonic diatoms, f – and Foraminifera, g – in bottom sediments from the Santa Barbara Basin, California (data from Lange, Burke, and Berger, 1990).

ing. And, if you believe in the greenhouse effect, this warming is likely to continue for some time to come. If this is so, then perhaps events such as the *Ceratium* increase and the recent occurrence of Atlantic indicators in the North Sea, have to be considered not simply as short-term noise, but as transient episodes forming part of an on-going trend, perhaps even as indicators of what may become the normal state of affairs, as possible

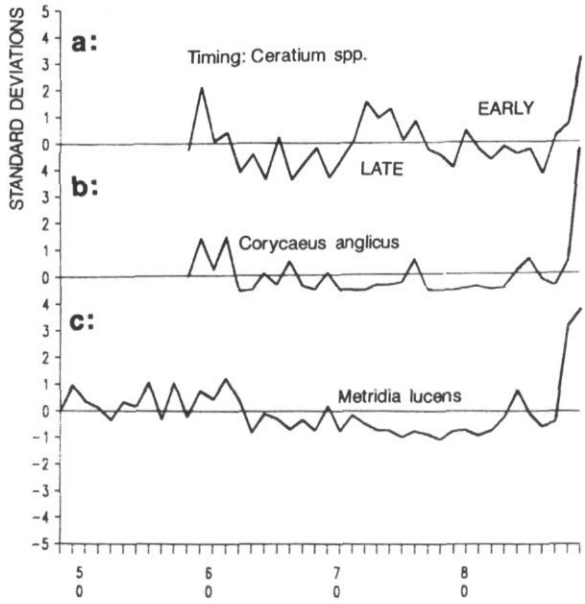

Figure 8. Standardized annual time series for the North Sea of a – the timing of the seasonal cycle for *Ceratium* spp., b and c – the abundance of the copepods *Corycaeus anglicus* and *Metridia lucens*.

pointers to the direction in which the underlying trends may go.

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Variability in phytoplankton biomass in the German Bight near Helgoland, 1980–1990

W. Hickel, J. Berg, and K. Treutner

Hickel, W., Berg, J., and Treutner, K. 1992. Variability in phytoplankton biomass in the German Bight near Helgoland, 1980–1990. – ICES mar. Sci. Symp., 195: 249–259.

The variability of total phytoplankton biomass and species composition was measured from surface-water samples taken every working day near the island of Helgoland, German Bight, North Sea. These data were analysed, together with data on inorganic nutrients, salinity, and Elbe river discharge. Main diatom growth periods were found in April and June through August, whereas non-diatom phytoplankton ("flagellates") bloomed in July and August. The biomass in the winter months consisted mainly of unidentified, small "μ-flagellates" of uncertain trophic state. Single peaks, or plankton blooms, varied considerably in size and time of occurrence within the vegetation period. Variations of single phytoplankton values ranged from almost nil to 1000 μg organic carbon/l during the year; variations of two orders of magnitude within a few days were not uncommon, reflecting sequences of different watermasses with their plankton populations. Monthly median values ranged from <10 to 350 μg C/l. While the total phytoplankton biomass near Helgoland increased during the last three decades, the potential eutrophication effect on phytoplankton is not very clear. For the decade 1980–1990, there is no apparent trend in phytoplankton increase or decrease, in contrast to other findings in the open North Sea. Analysing two periods of large Elbe river floods (1981/1982 and 1987/1988), an influence of additional nutrient supply on phytoplankton stocks of the inner German Bight was different: whereas late summer floods in 1981 resulted in very large phytoplankton stocks also near Helgoland, early (April) floods in 1987 and 1988 had no such consequences. From the hydrographical structure of the German Bight it is more likely that large river water quantities influence phytoplankton stocks first by increased density stratification and then by adding nutrients; this might be true particularly for the inner German Bight, whereas towards the outer German Bight, where nutrient concentrations rapidly decrease, the eutrophication effect might become more significant.

W. Hickel, J. Berg, and K. Treutner: Biologische Anstalt Helgoland, Notkestrasse 31, 2000–Hamburg 52, Germany.

Introduction

The variability of total phytoplankton biomass and species composition was measured from surface-water samples, taken every workday at Helgoland Roads, German Bight, North Sea (Fig. 8). These time-series measurements were started in 1962 (by Gillbricht) and comprise other biological parameters as well as hydrographic and inorganic nutrient measurements.

The sampling station, in a narrow channel between the two islands of Helgoland, represents a mixed water from the surrounding sea area and is considered to be representative for a larger area of this transition zone between estuarine and coastal water, and open North Sea water.

The last decade being the topic of the ICES Variability Symposium, the phytoplankton data from 1980–1990 are given here in more detail, together with inorganic nutrient and salinity data as well as Elbe river discharge volumes, which influence the German Bight in many ways.

The Helgoland Roads data have been analysed by Gillbricht (1983) for 1981 and Gillbricht (1988) for the period 1962–1986. Radach and Berg (1986), Radach *et al.* (1990), and Radach and Bohle-Carbonell (1990) analysed the Helgoland data from 1962 through 1984 in respect of the structure of the variance as well as to identify possible trends.

Methods

Sampling was done on Helgoland Roads (54°11.3′N 7°54.0′E) from a boat at a fixed time every workday,

thus not taking into account the tidal phase. The water column at the sampling station is about 5 m deep and always well mixed due to strong tidal currents. Therefore, the surface sample was considered to be representative for the water column.

The microplankton was counted with the inverted microscope after fixation with Lugol's iodine solution. The determination to the species level was only partly possible with the fixed material. More often, only the genera could be determined, or higher taxonomic levels. In the case of minute, naked μ-flagellates, only size classification was possible with the fast counting procedure applied here. Counted plankton was converted to organic carbon using factors calculated from size and shape of the plankters based on Hagmeier (1961). These biomass values, as organic carbon, were then summed up for diatom and non-diatom (="flagellate") biomass at least. The "flagellates" thus included various taxonomic groups; the dinoflagellates (without the heterotrophic *Noctiluca*) dominated by far in terms of biomass during their growth period (Fig. 1), though μ-flagellates were present in large numbers. In winter, however, the "flagellate" biomass consisted mainly of tiny, unidentified "μ-flagellates".

Salinity was measured with an inductive salinometer, and inorganic nutrients using the methods described in detail by Grasshoff *et al.* (1983). Water samples were analysed for nutrients immediately after sampling, except in the case of silicate which was analysed from deep-frozen samples. The Elbe river discharge data were taken from the yearly reports of the "Arbeitsgemeinschaft für die Reinhaltung der Elbe" (ARGE Elbe 1990a, b, c). The salinity and temperature data from the RV "Gauss" cruise in August, 1981, were measured with a bathysonde by members of the German Hydrographic Institute, Hamburg, and the nutrient analyses were made by Dr K. Eberlein, University of Hamburg.

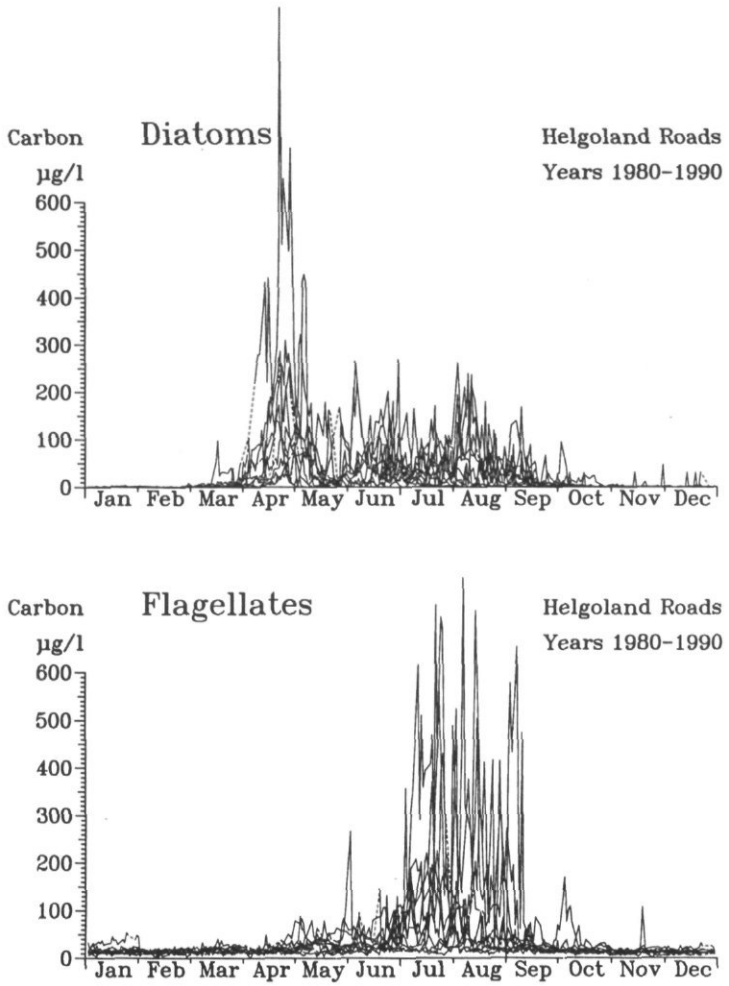

Figure 1. Annual cycles of diatoms and flagellates, Helgoland Roads (1980–1990). All year cycles projected into one year.

Results

Growth period of the phytoplankton

The growth period of diatoms and non-diatoms (="flagellates") was first determined by projecting all year-cycles 1980–1990 of the calculated biomasses into one year. Figure 1 shows the year-cycles of the single years. Figure 2 gives a scatter plot with the median, upper, and lower quartiles. Main diatom growth periods are obvious from the end of March, with peaks in April and May, but further diatom growth also occurred during the whole summer, until mid-September. Non-diatom phytoplankton ("flagellates"), however, are abundant in a much more restricted period, from June to mid-September only, with a clear peak in July. Obviously, the flagellates bloom under typical summer conditions with high temperatures and high regeneration rates of phosphate and ammonia, whereas diatom blooms are found both under "new" and "regenerated" nutrient conditions.

The biomass in the winter months consisted mainly of unidentified, small, naked "μ-flagellates" which might have been heterotrophic. Apart from them, very low phytoplankton biomasses were found in winter. Heterotrophic flagellates, frequent among the size-classes counted as μ-flagellates, would have no problems surviving in winter during poor light conditions, when autotrophic phytoplankters could thrive only in the surface water layer, and only during calm wind periods. Seston concentrations being high, and wind turbulence normally considerable, conditions for a net production of phytoplankton in the water column outside the vegetation period (March to September) would be rare and of short duration. This makes it unlikely that a considerable and constant autotrophic μ-flagellate population could possibly persist throughout the winter. Hence, the winter flagellate population, representing a biomass of about 10–15 μg C/l in the decade considered here, is likely to be heterotrophic at least during the winter period. The separate analysis of the phytoplankton during the vegetation period seems to be more appropriate.

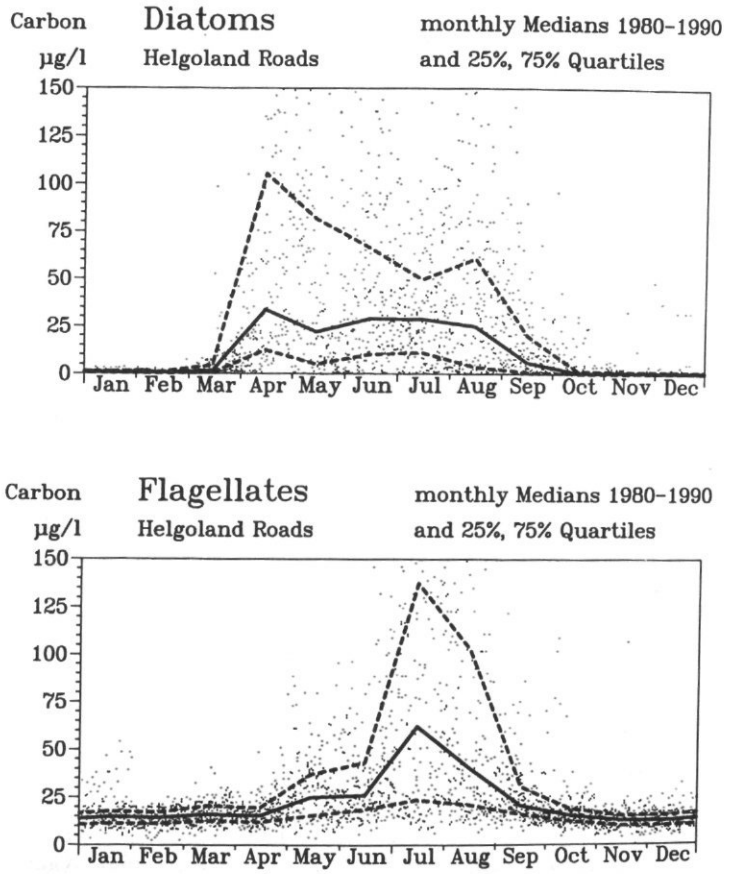

Figure 2. Variations in diatom and flagellate biomass at Helgoland Roads. All data for 1980–1990 pooled, with median, upper (75%), and lower (25%) quartiles.

Variation of phytoplankton stocks in time

Considering single values, phytoplankton concentrations varied considerably in size and time of occurrence within the vegetation period. Variations in biomass values ranged from almost nil to 1000 μg organic carbon/l. Variations of two orders of magnitude within a few days were not uncommon, reflecting sequences of watermasses with different plankton populations. Monthly median values ranged from <10 to 350 μg C/l (Fig. 6).

Considering yearly median biomass values of the vegetation periods, when a net production of the autotrophic plankton in the water column is possible, an increase of phytoplankton biomass for the last three decades seems evident (Fig. 3). Separating the potential bloom periods of diatoms and flagellates (according to Figs. 1 and 2), diatom stocks seem to increase up to the early 1980s only, whereas average flagellate stocks seem to increase during the 1980s (Fig. 3). However, the year-to-year variation in diatom and flagellate biomass during the growing season is about one order of magnitude higher than a possible upward trend. Also, the monthly medians of phytoplankton, diatom, and flagellate biomasses of the decade considered here show no clear trend (Fig. 6). The years with exceptional flagellate blooms are the prominent features. The large diatom stocks in 1985 were the result of peculiar hydrographic conditions: in 1984 and 1985, the coastal water zone was less extended due to low Elbe river water outflow (Fig. 7). In 1985, diatom blooms developed in the North Frisian coastal water (Bauerfeind *et al.*, 1990) in April and extended farther southward than usual, reaching Helgoland waters.

The long-term variations of the inorganic nutrients phosphate and nitrate are given in Figure 4 as yearly medians. This figure shows the rise and fall of phosphate concentrations as well as the late rise of nitrate concentrations. The rise of phosphate in the inner German Bight comes to an end around 1980, levelling off and soon declining. Nitrate concentrations, however, started to rise in the 1980s. The large Elbe river floods in 1987/1988 with their additional nitrogen loads (Gerlach, 1990; ARGE Elbe, 1990b) led to a marked increase in nitrate concentrations at Helgoland (Fig. 9); in the following years, too, these high levels remained in the coastal water.

This eutrophication process, first with phosphate, then with nitrate, leads to the question of possible biological effects. Though the absolute nutrient levels in the Helgoland waters are not growth-limiting for phytoplankton in summer (Gillbricht, 1988), the shift of N/P-ratios (Fig. 5) as the result of the above mentioned difference in P- and N-eutrophication might have had an effect on the composition of phytoplankton populations. Furthermore, there is always a steep gradient of nutrient concentrations from the German Bight coastal water body towards the open North Sea; the sites of potential growth-limiting nutrient concentrations in summer are not far away from Helgoland.

The influence of Elbe river freshwater runoff on phytoplankton stocks

The potential eutrophication effect on phytoplankton abundance is not very clear from the Helgoland data, certainly not in the decade considered here. A diatom increase might have taken place during the period of phosphate increase, whereas a flagellate increase seems to have occurred a decade later, along with a nitrogen increase. However, the year-to-year variability was an order of magnitude larger than a possible trend, and single events like the dinoflagellate blooms in 1981 tend to mask a trend.

If there was a direct influence of eutrophication by the river Elbe – as the main freshwater and nutrient source – on phytoplankton growth, then a correlation between the Elbe discharge volumes with phytoplankton stocks could be expected. Such relations could lead to more insight than calculating trends despite the large interannual variation. From Figure 7, two periods of particularly large Elbe river floods are obvious, in 1980/1982 and 1987/1988. The daily discharge values give the short-term peaks, whereas the lower graph of Figure 7 accumulates the discharge volumes for the months January through August of the respective years.

In 1987, the discharge volumes were particularly large. In the first half of the year, 21.7×10^9 m^3 of fresh water from the Elbe river had flowed into the German Bight, about the average discharge volume of a whole year. In 1987 and 1988, the major flood events took place in April, in contrast to 1981 when a late flood event was recorded in July/August (ARGE Elbe, 1990a, c).

These periods were evaluated here, as they could be expected to show an influence of additional nutrient supply on phytoplankton stocks of the inner German Bight. They were, on the other hand, examples of characteristic hydrographic situations caused by high freshwater inflow into the German Bight. In 1981, large Elbe river water inflow followed by a late (July/August) flood event resulted in a large area of the German Bight with a strong vertical stratification due to salinity gradients during a calm weather period. In spring 1987 and 1988, early (April) floods caused rather normally extended Elbe water plumes without such large stratified areas, due to windy weather.

The consequences of the river floods appearing at different seasons seemed to be strikingly different: whereas a late summer flood in 1981 resulted in the largest phytoplankton blooms found so far in the area, early (April) floods in 1987 and 1988 (Fig. 9) did not result in much larger phytoplankton stocks near Helgoland. This cannot be explained from additional inorganic nutrient supply alone; rather one would have

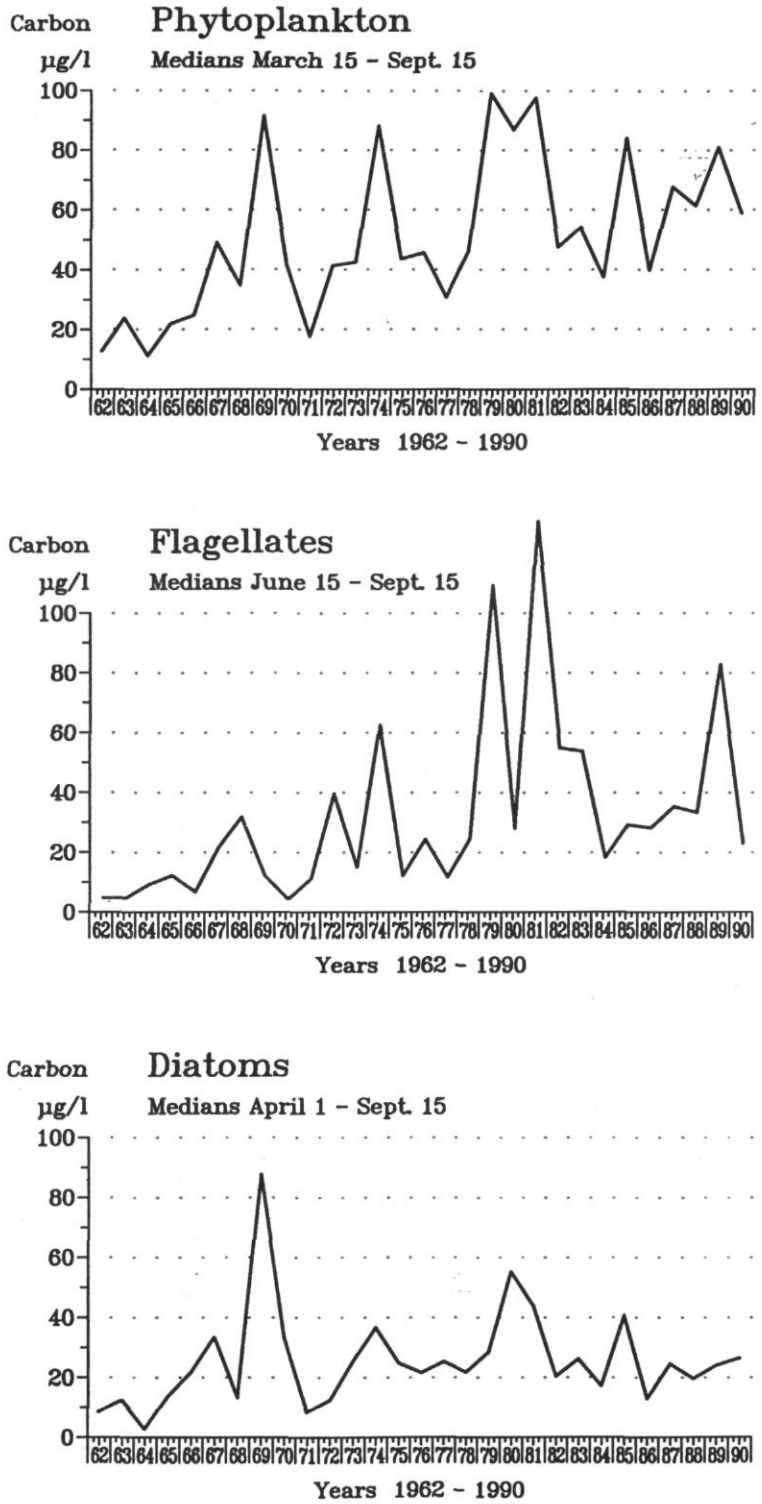

Figure 3. Phytoplankton biomass at Helgoland Roads (1962–1990). Yearly medians of the growing season (middle of March through mid-September, upper graph). Yearly medians of the potential blooming periods of diatoms (April through mid-September) and flagellates (mid-June through mid-September).

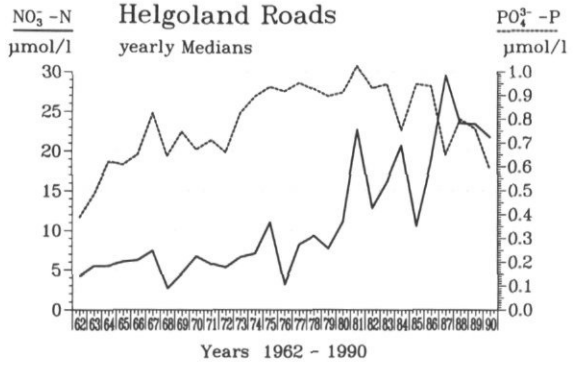

Figure 4. Nitrate and phosphate concentrations at Helgoland Roads (1962–1990), yearly medians.

expected the contrary. In 1987/1988, nitrate concentrations were twice as high, silicate concentrations almost ten times higher, as the average of the decade before. Only phosphate concentrations had decreased, indicating their origin from (constant) point sources and a dilution effect (Fig. 9). The particularly high nitrate concentrations resulted in a large nitrogen surplus in the coastal water even during summer (Fig. 5), when normally a phosphate surplus is found due to faster P-remineralization.

Additional data from the whole German Bight give more insight into the effects of Elbe river floods, namely their influence on vertical density stratification and the resulting growth conditions for phytoplankton. Figure 8 shows results from a RV "Gauss" cruise on 17–26 August 1981, just after the summer flood events of the Elbe river. On this cruise, a grid of stations as close as five nautical miles to each other was visited (10 miles for biological and chemical parameters). The Elbe water outflow and a long period of calm weather had caused an intensely stratified water mass (Fig. 8, as the density difference $\Delta\sigma t$ between the surface and bottom water samples).

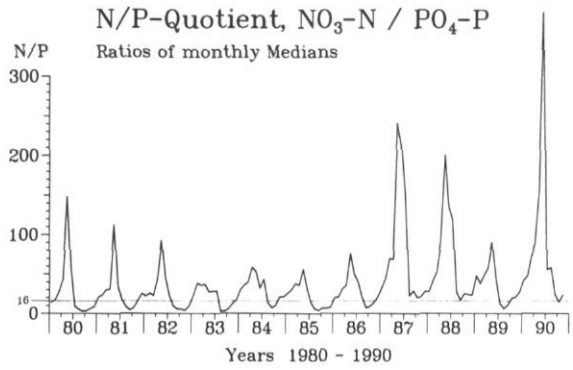

Figure 5. Nitrate-nitrogen to phosphate-phosphorus ratios at Helgoland Roads (1980–1990). Monthly median values.

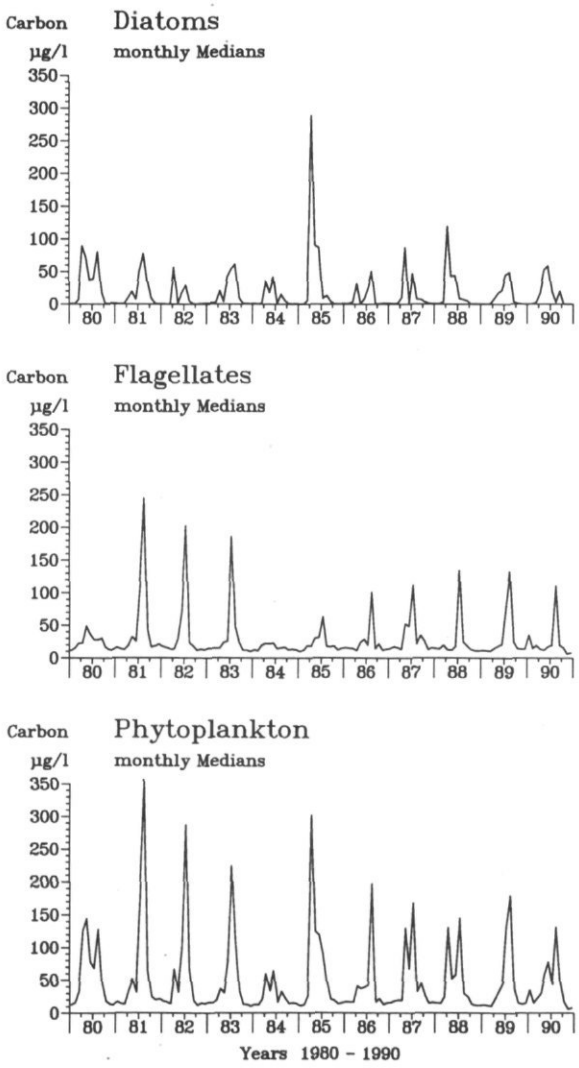

Figure 6. Phytoplankton, diatom, and flagellate biomass at Helgoland Roads (1980–1990), monthly medians.

At the edges of this stratified water body, extending in a northwest direction throughout the whole German Bight, large blooms of the dinoflagellate *Ceratium furca* were found. The southern bloom obviously had spread out as far as Helgoland (where it has been described by Gillbricht (1983); however, without taking into consideration its origin). In the stratified water, very low phosphate concentrations (<0.1 µmol/l) were found. As also silicate was depleted, a previous diatom bloom must have consumed the nutrients. Though the light conditions were favourable for plankton blooms within the stratified water, the nutrient exhaustion might have restricted the extension of the *Ceratium* blooms to the frontal zones.

In the southern *Ceratium* bloom area in August 1981 (Fig. 8), a highly significant negative correlation be-

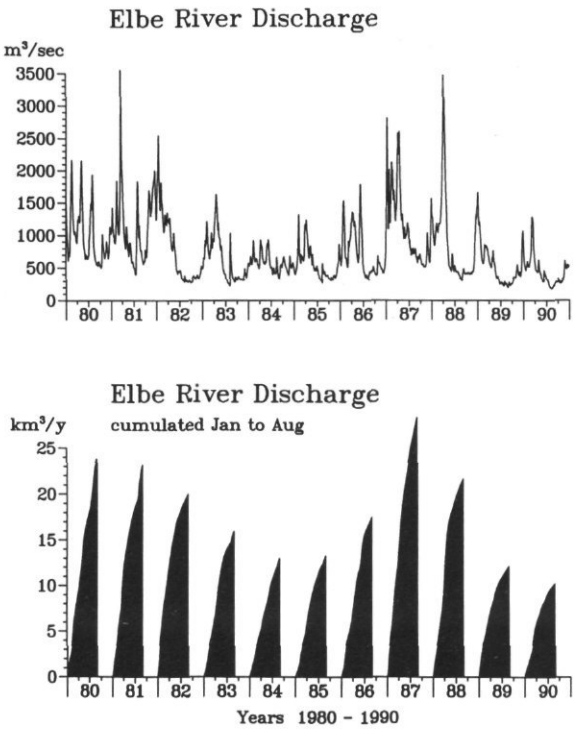

Figure 7. Elbe river discharge volumes (1980–1990). Single daily values (upper), cumulative volumes, January–August, lower graph.

tween *Ceratium* numbers and phosphate concentrations was found in the upper layer. This indicated that the *Ceratium* population was actively taking up phosphate. Therefore, the rising phosphate levels (and possibly other growth-enhancing substances) in the western frontal zone might have been the reason for the occurrence there.

Discussion

In the German Bight, phytoplankton blooms have often been found associated with hydrographic frontal systems (Krause *et al.*, 1986). The convergence zone of the German Bight (Goedecke, 1968) is a typical site of such blooms, and Figure 8 is a good example of blooms in frontal systems. Repeatedly, large phytoplankton blooms have also been found north of Helgoland within the north Frisian coastal water, e.g. diatom blooms in the years 1987 and 1988 (Hickel *et al.*, 1989; Bauerfeind *et al.*, 1990). Such blooms will not always be found by the Helgoland Roads sampling scheme, but they can drift to Helgoland during certain wind directions. From the hydrographical structure of the German Bight it can thus be understood why an additional eutrophication effect on phytoplankton need not be detected at the Helgoland monitoring station: the shallow, turbulent coastal water bodies off the river estuaries are too turbid to sustain large phytoplankton blooms, though nutrients are in good supply. Farther north and northwest, light conditions in the water column improve, mainly under the influence of salinity stratification and sedimentation of riverborne seston. Nutrient concentrations decline as phytoplankton production of the water column increases. Such patterns are known from other coastal waters directly influenced by turbid river water (Cadée, 1978) and can be detected by remote sensing from satellites (e.g. Szekielda, 1982; Aarup *et al.*, 1990).

To trace the point where phytoplankton growth starts in the German Bight, after the low salinity and bad light climate prevented marine phytoplankton growing in the Elbe estuary itself, the deviation of nutrient versus salinity plots from a linear relationship was analysed (Hickel, unpublished data from several cruises in the Helgoland Bight). During the phytoplankton growth season, this point where fast growing marine phytoplankton changes the nutrient/salinity regression by nutrient uptake was found mostly at a salinity between S = 28 and 29. This is the salinity at which Cadée (1978) found maximum production rates in the Zaire river plume and also from the Ems estuary. The S = 28 isohaline is normally found close to Helgoland. Until the bloom is fully developed, the northgoing residual current will have carried it to positions north of Helgoland.

The interpretation of plankton stocks in relation to nutrients from the Helgoland data is thus limited by the fact that the phytoplankton could originate from distant growing centres well away from Helgoland. The residual current, going normally from south to north (Hickel, 1980), can reverse in certain wind conditions, bringing watermasses from the northern German Bight with their larger plankton populations to Helgoland. These populations might have profited earlier from the eutrophication of the German Bight, but this might not be evident from Helgoland data.

Large freshwater quantities might support phytoplankton blooms first by increased density stratification, with the resulting improvement of light climate, and then by additional nutrient supply in the outer German Bight. To prove these effects, more horizontal surveys of the sort given in Figure 8 are needed, in addition to the continuing of the Helgoland time-series measurements. The latter should be continued by all means, as no other sampling scheme in a better suited area of the German Bight could possibly be conducted at such a high sampling frequency and low cost. It should now be self-evident that man-made changes versus the notoriously high natural variability in the inner German Bight can only be separated by time-series measurements of this kind.

Comparing the Helgoland time-series data with other long-term plankton records, the continuous plankton recorder (CPR) data might first be considered. Recent changes in the summer plankton of the North Sea

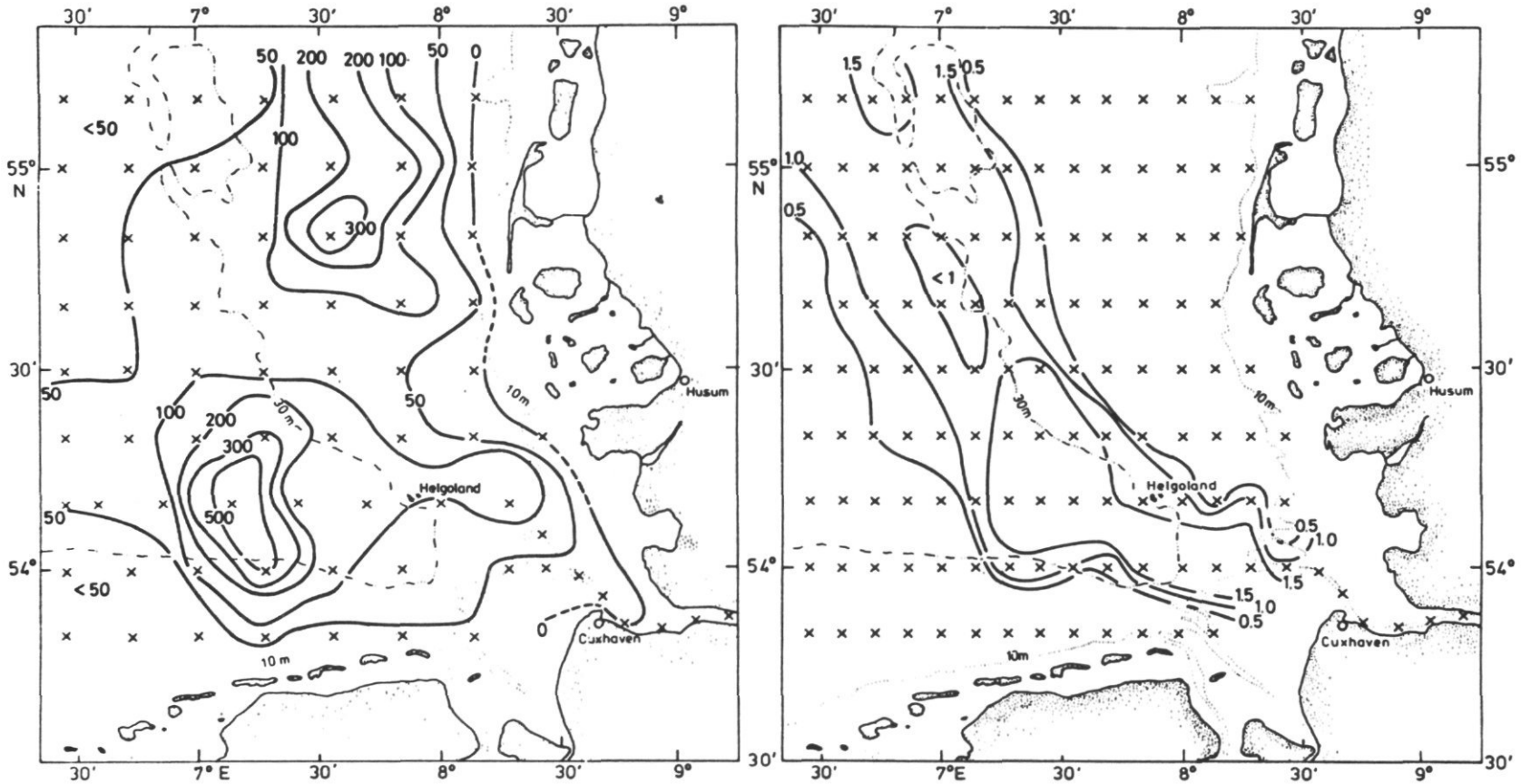

Figure 8. Density stratification (right) and *Ceratium furca* concentrations (left) in the German Bight in August, 1981 (following an Elbe river flood event in July/August). The water density stratification is given as the density difference $\Delta\sigma t$ between the surface and the bottom water; *Ceratium* numbers are mean values (in 10^3 cells/l) of the upper 10 m layer (three sample depths).

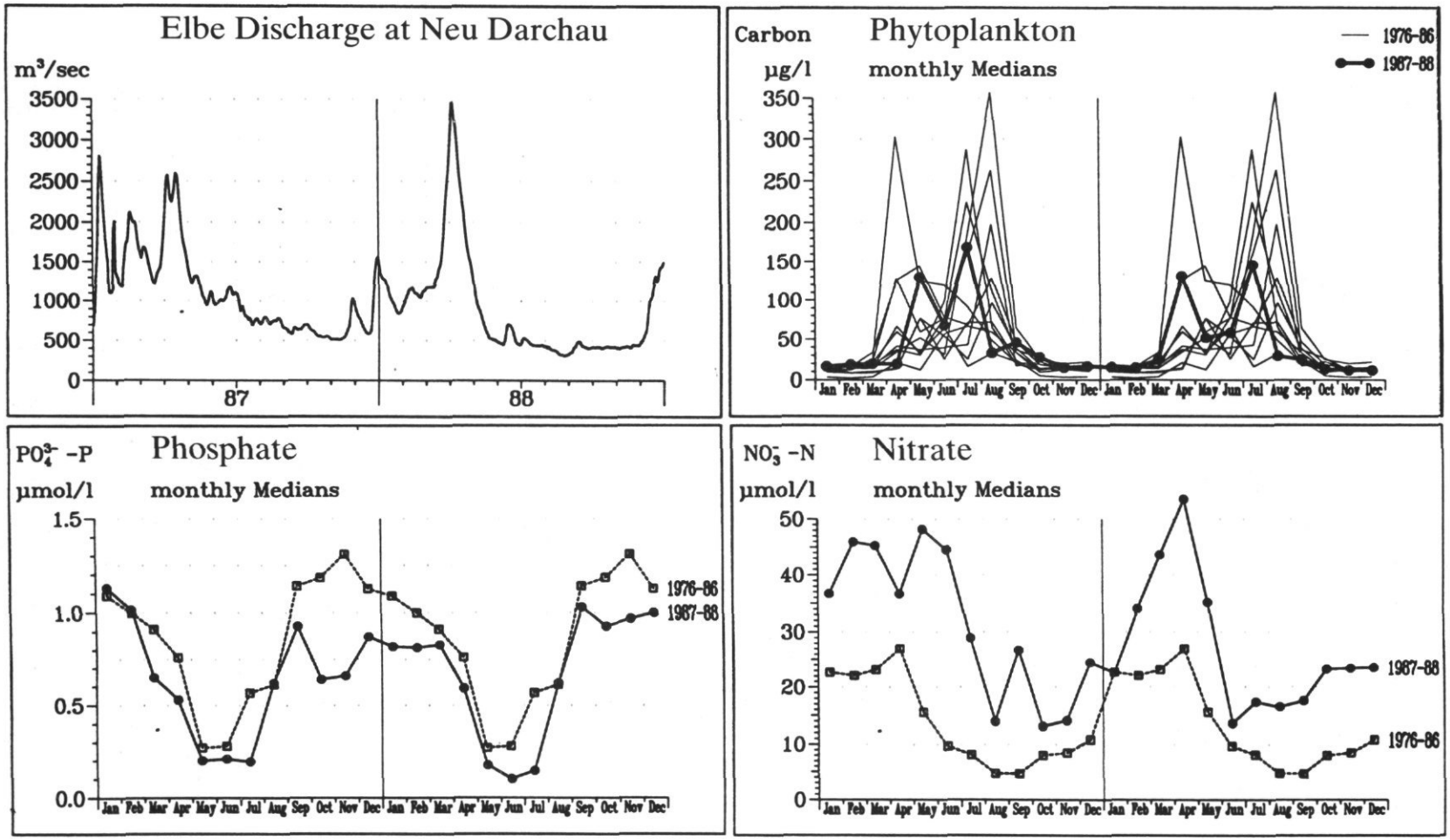

Figure 9. Elbe river water discharge at Neu Darchau, nutrients and phytoplankton at Helgoland in 1987 and 1988, as compared with long-term average values.

(Dickson *et al.*, 1991) comprised largely increased *Ceratium* abundance in the north-central North Sea along with increased stratification of the water column. The preference of *Ceratium* spp. for stratified water is well known. A westward extension of the Jutland coastal water may have supported the observed smaller increase of *Ceratium* spp. during the late 1980s in the German Bight, according to these authors. The Helgoland data failed to show this increase, or an earlier occurrence of dinoflagellate blooms in the year-cycle as found in the open North Sea, though the salinity decrease and nutrient increase by the Elbe river floods in 1987 and 1988 were well documented. Only the 1981 blooms after the unusual extension of a stratified water zone in the German Bight confirm the views of Dickson *et al.*

The extensive literature on phytoplankton changes in the North Sea on the basis of CPR data (e.g. Radach, 1984 for a review) does not cover most of the coastal water of the North Sea adequately. This narrow and eutrophic water body off the Dutch and German coasts is separated from the open North Sea by frontal systems (see Otto *et al.*, 1990, for a review), and a response of phytoplankton populations different to those in the open North Sea should not be surprising. Long-term changes within this coastal water are monitored by the Marsdiep time series (Cadée, 1988, 1991), besides the Helgoland Roads time series.

In the westernmost Dutch Wadden Sea, the Marsdiep area, the Rhine water influence is a more stable and predictable factor than the Elbe river influence is for Helgoland. The Rhine water keeps close to the Dutch coast, where its quantities are positively correlated with phytoplankton biomass (Gieskes and Schaub, 1990). The increase in phytoplankton in the Marsdiep area has been described in many papers by Cadée, both as increased primary production (Cadée, 1986) and phytoplankton abundance (Cadée, 1991), but this increase was due to non-diatom plankton, particularly *Phaeocystis globosa*. This species is rarely a major component of the Helgoland phytoplankton biomass, though it regularly occurs there. The main blooms of *Phaeocystis* are found more in the east Frisian and north Frisian coastal waters close to the coast.

The Elbe river water, discharging only about one-third of the Rhine river volume, is at normal discharge rates transported to the North between Helgoland and the peninsula of Eiderstedt, but after high river discharge or strong easterly winds, the Elbe river water plume will reach Helgoland. On the other hand, it has been mentioned before that occasionally reversed residual surface current flow can carry plankton bloom patches from the northern German Bight to Helgoland, adding to the variability of plankton time series at the Helgoland Roads station.

The review of Reid *et al.* (1990) shows the complexity of phytoplankton dynamics in the North Sea. The plankton distribution is even more complex in time and space in the German Bight, where recurrent patterns are hard to recognize. To improve this knowledge, and to relate the German Bight data with open North Sea data as documented by the CPR, the Helgoland plankton data will be reevaluated in more detail (e.g. for *Ceratium* species), together with the numerous (partly unpublished) data on horizontal quasisynoptic surveys of the whole German Bight.

Among the much discussed potential effects of eutrophication are the oxygen depletion events 1981–1983 in bottom water of the southeastern North Sea associated with mass mortality of benthic animals (von Westernhagen and Dethlefsen, 1983; von Westernhagen *et al.*, 1986). The possible sources and the effects on the bottom fauna were subsequently investigated (Hickel *et al.*, 1989). Besides the obviously dominating role of water column stratification during calm weather periods, a contribution of the large phytoplankton blooms of these years (Fig. 6) to the organic detritus load and thus oxygen consumption in the bottom water would be likely; however, the data from the respective years were not sufficient to prove this in a particular case. Only in 1989 were low oxygen concentrations again found in the bottom water of the northern German Bight (north of Sylt) and the Danish coastal waters of Horns Rev (Niermann, 1990); this might have been the result of plankton blooms which escaped detection by the Helgoland Roads sampling scheme.

Acknowledgments

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ICES mar. Sci. Symp., 195: 260–267. 1992

Variabilité hydrobiologique et productivité secondaire pélagique: une approche par la mesure de l'activité de l'aspartate transcarbamylase

Jean-Pierre Bergeron

Bergeron, J-P. 1992. Variabilité hydrobiologique et productivité secondaire pélagique: une approche par la mesure de l'activité de l'aspartate transcarbamylase. – ICES mar. Sci. Symp., 195: 260–267.

La productivité secondaire pélagique est estimée par la mesure de l'activité d'une enzyme, l'aspartate transcarbamylase (ATC), d'échantillons du peuplement mésozooplanctonique total. La méthode a été mise en oeuvre au cours de trois périodes printanières consécutives dans le sud du bassin occidental de la Manche. L'activité de l'ATC et la biomasse protéique du mésozooplancton sont liées par une relation de type allométrique, dont les paramètres peuvent prendre différentes valeurs. La variabilité spatio-temporelle des activités de l'ATC associée aux conditions de l'environnement hydrobiologique est illustrée par divers exemples. On peut ainsi mettre en évidence (1) une évolution saisonnière normale se déroulant selon un processus harmonieux, (2) une variabilité spatiale "instantanée", suivie d'une évolution temporelle singulière, (3) l'apparition d'une perturbation de l'écosystème. La cohérence globale des résultats renforce l'hypothèse de l'existence de propriétés systémiques au niveau de perception adopté. L'activité de l'ATC apparaît comme l'expression de l'intégration biologique par le système mésozooplanctonique des influences complexes des multiples facteurs de l'environnement.

Jean-Pierre Bergeron: IFREMER, Direction des Ressources Vivantes, Laboratoire OCEHAL, BP 1049, 44037 Nantes Cedex 01, France.

Introduction

Empreinte des fluctuations de l'environnement physique, la variabilité hydrobiologique du milieu marin exerce, à des degrés divers et selon des modalités multiples, une influence sur tous les composants de la biocénose. De ceux-ci, le peuplement mésozooplanctonique constitue un premier maillon essentiel, permanent et omniprésent dans l'écosystème pélagique, de métazoaires hétérotrophes subissant immédiatement les effets de cette variabilité. Ses caractéristiques structurales et fonctionnelles, ainsi que le métabolisme global qui en émerge, sont par essence la résultante d'une intégration biologique de l'ensemble des facteurs de l'environnement.

Une nouvelle méthode d'approche de la production secondaire pélagique a été proposée (Bergeron, 1983), fondée sur la mesure de l'activité d'une enzyme, l'aspartate transcarbamylase (ATC), d'échantillons du peuplement mésozooplanctonique total. Une première mise en oeuvre a été tentée au cours de trois périodes printanières consécutives, de 1978 à 1980 (Tableau 1), dans le sud du bassin occidental de la Manche (Fig. 1), selon les méthodes décrites précédemment (Bergeron, 1986). Les principaux résultats issus de ces travaux ont déjà été présentés (Bergeron, 1983, 1986, 1990); cependant, les relations entre les variations de l'activité de l'ATC et les conditions de l'environnement hydrobiologique n'ont été que très succinctement évoquées pour étayer la cohérence de ces premiers résultats.

Tableau 1. Dates d'exécution des campagnes.

Suroît I	30 mars	au	4 avr. 1978
Suroît III	14 avr	au	17 avr. 1978
Ecomanche I	12 mai	et	13 mai 1978
Thalia 5	20 mars	au	23 mars 1979
Thalia 6	03 avr	au	06 avr. 1979
Thalia 7	02 mai	au	05 mai 1979
Thalia 8	19 juin	au	22 juin 1979
Thalia 10	25 mars	au	29 mars 1980
Thalia 11	08 avr	au	12 avr. 1980
Thalia 12	05 mai	au	10 mai 1980
Thalia 13	19 juin	au	23 juin 1980

Il apparaît maintenant opportun de réexaminer cet aspect de façon plus détaillée, notamment afin d'évaluer l'aptitude de la méthode à fournir un niveau de perception pertinent pour une estimation globale de la variabilité hydrobiologique et de ses effets sur le système mésozooplanctonique.

Contexte océanographique

La zone d'étude s'insère dans le vaste ensemble de mer épicontinentale du nord-ouest de l'Europe, où le transfert à long terme des masses d'eau subit l'influence de la dérive nord-atlantique, extension faible et diffuse du Gulf Stream: c'est la composante quasi stationnaire des écoulements aux frontières. A l'échelle régionale de la Manche, c'est le régime macrotidal qui domine largement la circulation résiduelle (Salomon et Breton, 1991), qui ne peut être altérée que temporairement par les vents ou localement par des gradients de densité, ces derniers étant pratiquement inexistants dans le secteur auquel on s'intéresse ici. Le modèle mathématique proposé par Salomon et Breton (1991) fait apparaître que, sous l'effet de la seule marée, l'alimentation de la Manche par les eaux d'origine atlantique s'effectue à proximité immédiate des côtes de la Bretagne, que le flux longe avant de s'orienter vers le nord du bassin occidental. La zone d'étude est donc le siège d'une circulation résiduelle d'ouest en est, de faible vitesse (de l'ordre de 2 cm s^{-1}) si l'on néglige l'interaction possible du vent. Des estimations expérimentales de diverses natures (citées dans Salomon et Breton, 1991) suggèrent qu'en règle générale le transport global d'eau de l'Atlantique vers la Mer du Nord est sensiblement plus important, ce qui peut être vraisemblablement attribué à un renforcement du phénomène par les vents dominants de cette région. Ce schéma général de circulation, d'ailleurs connu dans ses grandes lignes depuis plusieurs décennies, est amplement conforté par des études de dispersion de radionucléides artificiels (Germain *et al.*, 1986; Guegueniat *et al.*, 1986) et à l'évidence détermine la structure hydrologique standard de la Manche (Fig. 2 pour le mois d'avril), que révèlent les cartes de répartition des valeurs moyennes, établies sur 50 ans, de température et salinité mesurées en surface (CIEM, 1962). Il mérite d'être souligné enfin que, du fait de la modeste profondeur et grâce aux turbulences engendrées par les marées et les vents, les eaux dans ce secteur sont bien mélangées et la plupart du temps très homogènes du fond à la surface.

Variabilité spatio-temporelle des activités de l'ATC en relation avec les conditions de l'environnement hydrobiologique

Evolution saisonnière harmonieuse en 1979

Hydrobiologie

Dans la zone d'étude en période printanière, les isohalines sont généralement orientées sud-ouest nord-est, les températures montrent un gradient décroissant de

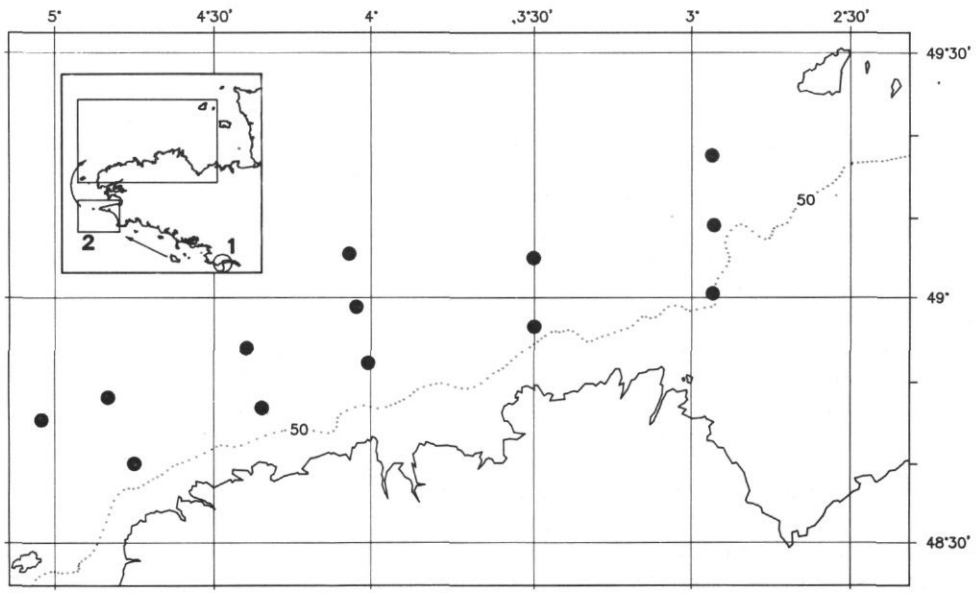

Figure 1. Zone d'étude et positions des stations d'échantillonnage en 1979 et 1980 (réseaux différents en 1978, cf. Fig. 7). En cartouche sont indiquées les situations respectives de la zone d'étude, de l'estuaire de la Loire (1) et du secteur d'échantillonnage particulier à l'année 1980 (2) correspondant au cartouche de la figure 9.

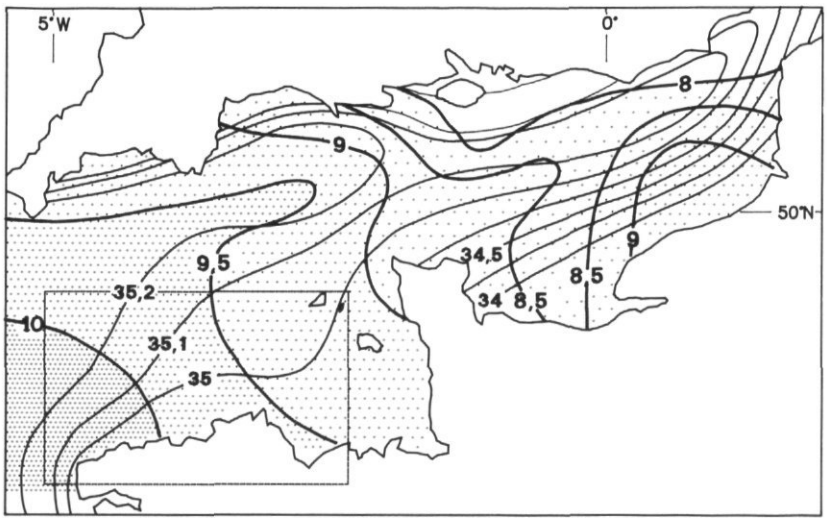

Figure 2. Situation hydrologique standard en avril pour l'ensemble de la Manche: répartition des valeurs moyennes mensuelles (1905–1954) de température et de salinité relevées en surface (d'après CIEM, 1962). Le cadre rectangulaire délimite la zone d'étude.

l'ouest vers l'est (CIEM, 1962; Fig. 2). En 1979, bien que la température de l'eau fût inférieure à la normale en mars à la suite d'un hiver particulièrement rigoureux, la structure hydrologique de la zone était conforme à ce schéma. Les quatre campagnes ont donné l'image d'une évolution harmonieuse de l'environnement hydrobiologique, avec notamment augmentations régulières et continues de la température, de la salinité et des teneurs en chlorophylle a, au cours de toute la période printanière.

Activité ATC

L'activité d'une enzyme s'exprime classiquement par rapport à une référence pondérale, le mg de protéines en général: c'est l'activité spécifique de l'enzyme (ici ATCs, en u ATC mg prot.$^{-1}$), qui permet la comparaison simple et immédiate de plusieurs échantillons. Mais au niveau de perception adopté, à l'instar de nombreux taux physiologiques ou fonctions métaboliques mesurés sur des organes ou des organismes, l'activité de l'ATC d'un peuplement mésozooplanctonique présente en outre une propriété fondamentale: ses variations spatio-temporelles (en valeurs brutes, exprimées par unité de volume d'eau échantillonnée: u ATC m^{-3}) sont le plus souvent liées à celles de la biomasse protéique (P) par une relation de type allométrique, dont les paramètres peuvent prendre différentes valeurs. La relation s'écrit ainsi: ATC = β P^{a}. L'activité spécifique (ATCs) équivaut au rapport des deux variables: ATCs = ATC/P = β P^{a-1}.

En 1979, les valeurs des activités spécifiques de l'ATC sont relativement homogènes sur l'ensemble de la zone. De plus, l'évolution au cours de la période printanière apparaît harmonieuse et ses caractéristiques principales présentent de remarquables analogies avec celles que l'on observe chez des organismes en équilibre stable avec leur milieu (Bergeron, 1982), ce qui a déjà été souligné et analysé précédemment (Bergeron, 1983): très classiquement, le segment de droite a (Fig. 3) traduit en effet une augmentation préalable de l'activité spécifique (pente supérieure à 1), qui est suivie d'une diminution progressive (segment b de pente inférieure à 1), représentative de l'essentiel de la maturation du peuplement.

Variabilité spatiale "instantanée" et évolution temporelle singulière en 1978

Au mois d'avril 1978, la structure hydrologique de la zone et son évolution entre les deux premières campagnes (Aminot et Kerouel, 1978) présente des particularités remarquables, dont il est aisé de définir quels sont les déterminants physiques majeurs, liés aux conditions météorologiques. Le régime des vents mérite d'être considéré en premier lieu: la région subit généralement un flux dominant d'ouest, ce qui fut le cas en 1979 (Fig. 4); mais la période précédant la campagne Suroît I fut marquée par un renforcement très net de ces vents soufflant en tempête et accompagnés de fortes précipitations; lors de Suroît I, on assista à un renversement brusque des vents qui s'établirent de secteur est-nord-est et persistèrent ainsi, relativement forts, durant plusieurs jours entre les deux campagnes (Fig. 4). Deuxièmement, la partie centrale de la côte nord de la Bretagne présente une double échancrure (entre 3°50′ et 4°W, cf. Fig. 1), ce sont les estuaires des deux principales rivières se jetant dans cette région; l'apport d'eau continentale

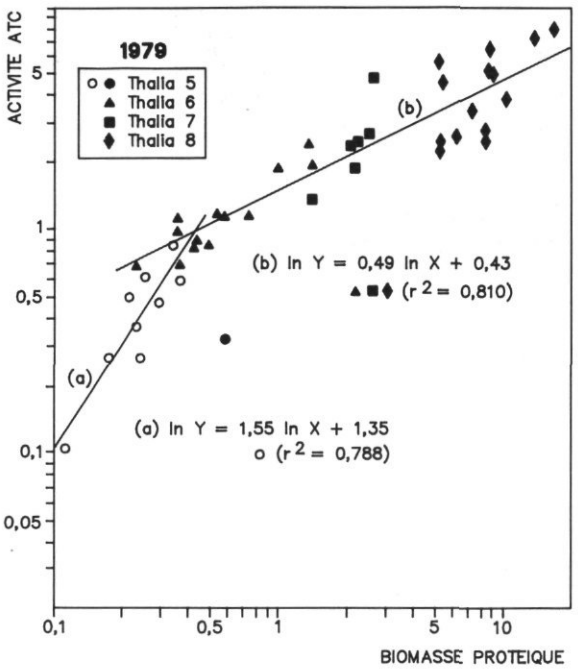

Figure 3. Variations en coordonnées ln-ln (népérien) de l'activité de l'ATC (exprimée en u ATC) en fonction de la biomasse protéique (exprimée en mg) au cours de la période printanière de 1979. Les deux variables sont rapportées au m^3 d'eau échantillonnée. Les calculs des paramètres des droites de régression et des coefficients de corrélation ont été effectués à l'aide des couples de valeurs représentés par les symboles figurant respectivement auprès de chacune des équations.

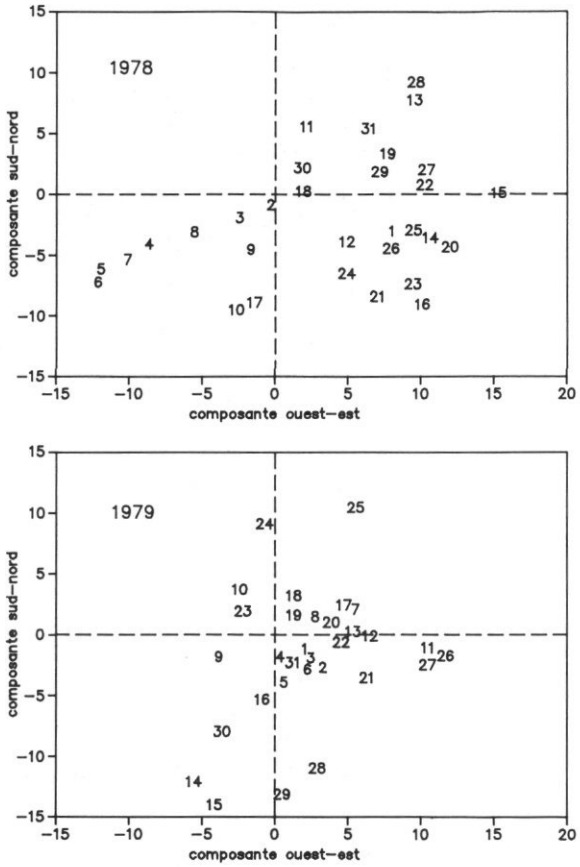

Figure 4. Régimes quotidiens des vents pour les périodes du 11 mars au 10 avril des années 1978 et 1979. Chaque nombre correspond à la date, sa position indique le sens et la vitesse (exprimée en m s^{-1}) du vent observé au sémaphore de l'île d'Ouessant (C. Leroy, comm. pers.: programme Climapêche, d'après fichier des sémaphores, Météo France).

demeure ici généralement modeste, mais, à la suite de pluies abondantes telles que celles qui viennent d'être signalées, une augmentation du débit de ces rivières peut survenir et se traduire par une dessalure locale accompagnée d'un enrichissement en éléments nutritifs. Enfin, l'hiver 1978, bien qu'à un degré moindre que celui de 1979, fut un hiver froid comme en témoignent les anomalies négatives des valeurs de température de l'eau (C. Leroy, comm. pers.: programme Climapêche, d'après fichier SSTGASC, Météo France).

Hydrobiologie

Sous l'influence directe de ces conditions, une structure hydrobiologique très particulière apparaît lors de la campagne Suroît I (Fig. 5), où trois masses d'eau principales peuvent être individualisées. Anormal si l'on se réfère à la situation standard (Fig. 2), un gradient thermique très marqué (0.8 à 1°C en moins de 15 milles) situé entre 3° et 3°30'W isole à l'est une masse d'eau froide (température inférieure à 8.5°C), où l'on relève de faibles teneurs en chlorophylle a, inférieures à 0.8 μg l^{-1} (Fig. 6). A l'ouest de ce front, dans une eau dont la température est comprise entre 9 et 10°C, on distingue un secteur sous influence océanique, révélé par des valeurs de salinité supérieures à 34.8–34.9 et où les teneurs en chlorophylle a sont généralement comprises entre 1 et 2 μg l^{-1}, et un secteur de plus faible salinité influencé par les apports continentaux et caractérisé par des teneurs en chlorophylle a supérieures à 2.5 et atteignant 3.7 μg l^{-1}. L'eau froide de l'est de la zone est vraisemblablement l'empreinte persistante de l'hiver froid précédent. Le resserrement des isothermes et la forme très particulière des isohalines, notamment dans le sud-ouest de la zone, traduisent bien le forcing exercé par les vents d'ouest ayant régné durant les deux décades qui ont précédé la campagne Suroît I.

Le brusque renversement, évoqué antérieurement, du sens des vents dominants est certainement à l'origine de l'important refroidissement des eaux de la portion ouest se manifestant ensuite à la mi-avril, lors de la campagne Suroît III (Fig. 5). On observe de plus une homogénéisation globale de la zone, accompagnée d'une sensible diminution des teneurs en chlorophylle a (Fig. 6). Cependant l'isotherme 8.5°C est apparemment

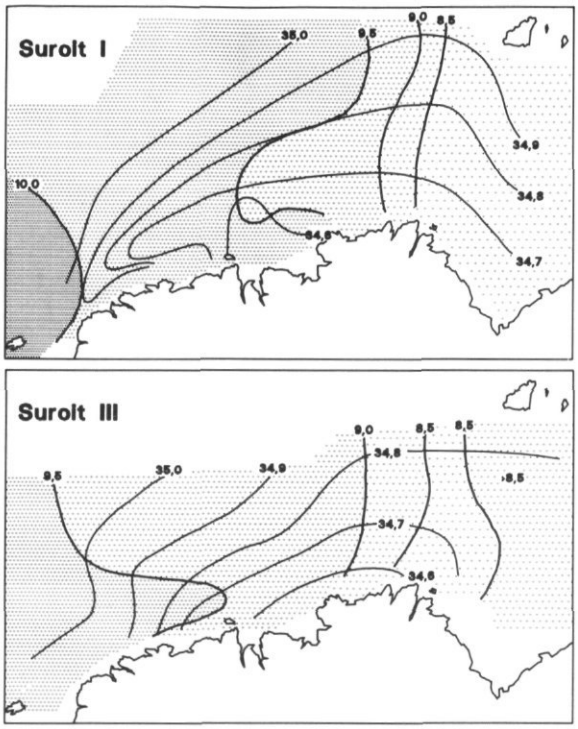

Figure 5. Répartition des valeurs de température et de salinité relevées en surface au printemps 1978 (d'après Aminot et Kerouel, 1978).

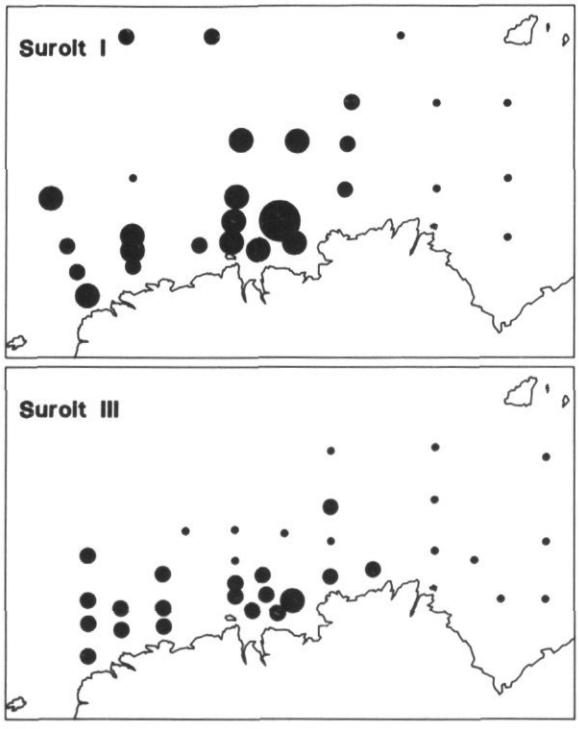

Figure 6. Répartition des teneurs en chlorophylle a (exprimées en $\mu g\ l^{-1}$) dans la couche superficielle au printemps 1978 (d'après Aminot et Kerouel, 1978). Les quatre tailles croissantes des symboles représentent respectivement les classes de valeurs 0–1, 1–2, 2–3, et 3–4.

demeurée stable, la portion est, probablement protégée du refroidissement sous l'influence des vents par le Cotentin, montre même un léger réchauffement et, malgré une apparente pauvreté en phytoplancton, une nette sursaturation en oxygène dissous témoignant une activité photosynthétique récente ou en cours (Aminot et Kerouel, 1978).

Activité ATC

Les activités spécifiques de l'ATC du peuplement mésozooplanctonique mesurées lors de la campagne Suroît I (Fig. 7) présentent une distribution spatiale tout à fait cohérente au regard de la structure hydrobiologique qui vient d'être décrite. On distingue en effet trois principaux groupes de valeurs: à l'est, où le processus de production printanière ne semble qu'a peine engagé, sont relevées les activités les plus faibles; à l'ouest, dans les eaux sous influence océanique se trouve un ensemble de valeurs d'une remarquable homogénéité et de niveau intermédiaire (valeur moyenne = 0.62; écart type = 0.09; pour 16 mesures); dans la zone plus côtière enfin, les activités les plus élevées sont associées aux eaux de plus faible salinité et les plus riches en phytoplancton. A ce stade de l'évolution de l'écosystème, la richesse de l'environnement trophique du mésozooplancton, traduite par l'abondance du phytoplancton, apparaît comme le déterminant majeur de la variabilité des activités de l'ATC puisque celles-ci sont significativement corrélées ($p < 0.01$) aux taux de chlorophylle a disponible par unité de biomasse protéique (Bergeron, 1986).

Ce n'est plus le cas deux semaines plus tard. lors de la campagne Suroît III, où la baisse des activités de l'ATC peut être logiquement imputée au refroidissement du milieu précédemment évoqué, probablement associé à une modification momentanée de l'advection. La persistance, le long de la côte à l'ouest, d'une petite fraction du peuplement présentant encore des valeurs plus élevées (Fig. 7) suggère que le processus est en cours de réalisation au moment de la campagne. Dans la portion est de la zone, malgré les faibles teneurs en chlorophylle a, le maintien d'activités spécifiques identiques à la majorité de celles qui étaient mesurées deux semaines auparavant est tout à fait conforme à l'analyse et aux conclusions d'Aminot et Kerouel (1978) concernant l'évolution entre les deux campagnes du processus de production phytoplanctonique, qui, selon ces auteurs, s'est progressivement déplacé vers l'est et dont une phase d'activité maximale s'est vraisemblablement produite juste avant la campagne Suroît III. On voit ici l'illustration d'une faculté capitale de la méthode: à ce niveau d'intégration

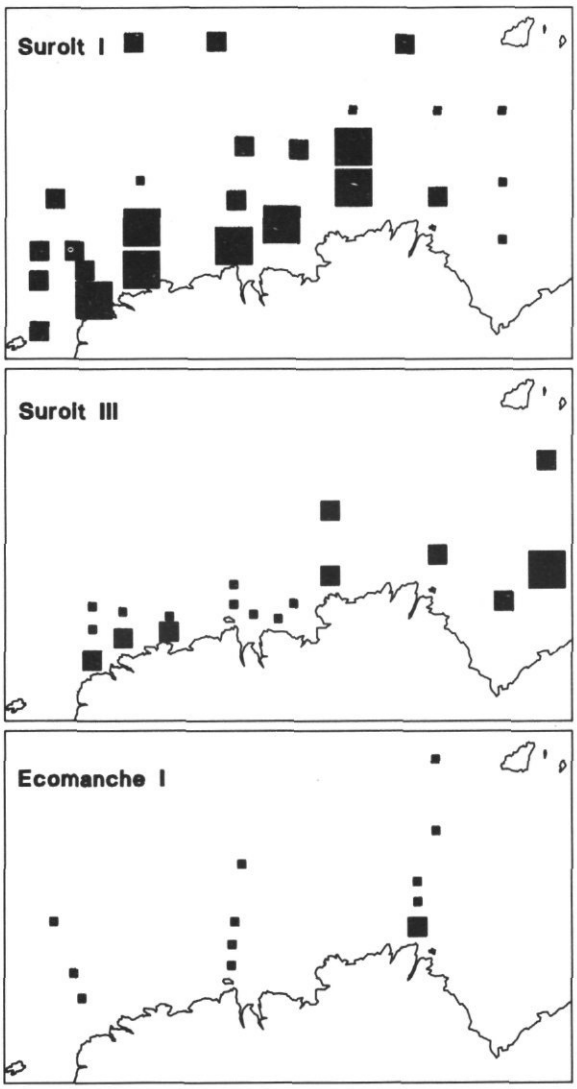

Figure 7. Evolution de la distribution spatiale des valeurs d'activité spécifique de l'ATC (exprimées en u ATC mg prot.$^{-1}$) au cours du printemps 1978. Les trois tailles croissantes de symboles représentent respectivement les classes de valeurs variant conformément aux relations e (ou $ATC_s < 0.3$), d et c (ou $ATC_s > 1$) de la figure 8.

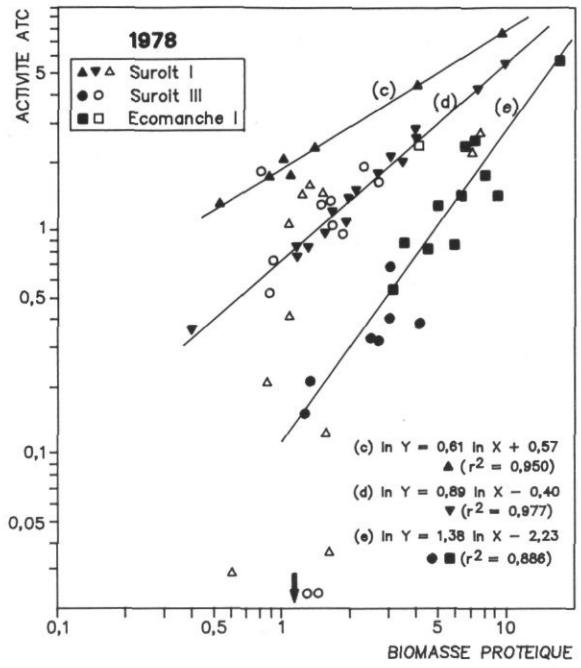

Figure 8. Variations en coordonnées ln-ln (népérien) de l'activité de l'ATC (exprimée en u ATC) en fonction de la biomasse protéique (exprimée en mg) au cours de la période printanière de 1978. Les deux variables sont rapportées au m^3 d'eau échantillonnée. Les calculs des paramètres des droites de régression et des coefficients de corrélation ont été effectués à l'aide des couples de valeurs représentés par les symboles figurant respectivement auprès de chacune des équations. On remarquera, pour la campagne Suroît I, un nombre d'observations supérieur à celui apparaissant sur la figure 7: la raison en est que certaines stations de la portion ouest ont été échantillonnées à deux reprises (cf. Bergeron, 1986).

biologique, le système conserve la mémoire d'événements influents passés, les mesures de l'activité de l'ATC sont susceptibles d'en révéler la résultante sur les mécanismes intimes impliqués dans l'élaboration de matière vivante.

Contrairement aux mesures effectuées en 1979, celles de 1978 ne permettent pas d'établir une relation allométrique unique (Fig. 8). A de très rares exceptions près, les variations des activités brutes de l'ATC sont étroitement associées à celles de la biomasse protéique par trois relations bien distinctes, correspondant aux trois grands niveaux d'activité spécifique mis en évidence (Fig. 7). L'intérêt fondamental de ce résultat est de montrer que les variations des activités de l'ATC des systèmes mésozooplanctoniques ne sont pas continues, mais discrètes, obéissant à des relations de type allométrique dont les paramètres semblent caractériser des entités fonctionnelles différentes et bien individualisées dans l'espace. En outre, ces entités semblent persister: les valeurs mesurées lors de Suroît III se conforment aux mêmes relations que celles établies pour Suroît I. Ainsi la campagne Ecomanche I, pour laquelle on ne dispose d'aucun descripteur de l'environnement hydrobiologique, semble indiquer que le système mésozooplanctonique habitant la zone est du même type, un mois plus tard, que celui qui fut échantillonné en un certain nombre de stations lors de la campagne Suroît III (Fig. 7 et 8). On peut sans doute raisonnablement en inférer que s'est produite entre temps une homogénéisation de la zone sans perturbation majeure du milieu.

Perturbation de l'écosystème en 1980

Hydrobiologie

Au début de la période printanière de 1980, la structure hydrobiologique de la zone était tout à fait conforme au schéma standard (Fig. 2). La seconde campagne, menée en avril, permit de révéler une dessalure (Fig. 9), surtout sensible au niveau des stations les plus occidentales, très probablement imputable à une incursion d'eau provenant de la Loire (P. Le Corre, comm. pers.); l'hypothèse de cette origine est renforcée par les valeurs de salinité relevées en des points situés en amont (Fig. 9) sur le parcours normal effectué par cette eau (flèches, Fig. 1). L'évolution de la valeur moyenne de la salinité, encore inférieure à 35 à la fin de juin, indique la persistance de la dessalure au cours de toute la période printanière (Fig. 9).

Activité ATC

Compte tenu de la remarquable étroitesse de la relation qui les associe aux variations de la biomasse protéique, les activités enzymatiques mesurées au cours des deux premières campagnes de 1980 semblent refléter une très grande homogénéité du peuplement mésozooplanctonique (Fig. 10). Au mois de mai apparaît une profonde perturbation de cette situation, révélée par la disparition totale de toute relation allométrique. Malgré le décalage temporel apparent entre les deux événements, il est difficile de ne pas attribuer cette perturbation à l'invasion de la zone par des eaux de salinité anormalement faible et de nature inhabituelle pour la région. La campagne Satir 3, exécutée à la mi-juin 1980, vient appuyer cette hypothèse en donnant une confirmation non seulement de la perturbation que connaît l'hydrologie normale du secteur, mais également des répercussions potentielles de celle-ci sur le fonctionnement du système mésozooplanctonique: pour des biomasses du même ordre de grandeur se manifeste une dispersion indentique des activités de l'ATC (Fig. 10). Dans la zone d'étude, celles-ci reprennent des valeurs plus élevées et une nouvelle relation allométrique se dessine à la fin de juin (Fig. 10), différente de celle établie pour les deux campagnes de mars et avril; la diminution des activités spécifiques exprimée par cette différence traduit vraisemblablement la persistance de l'empreinte de l'anomalie hydrobiologique sur le système mésozooplanctonique. Les arguments théoriques susceptibles de permettre une compréhension globale de l'enchaînement des divers aspects que présente la variabilité des activités de l'ATC durant cette période ont été exposés antérieurement (Bergeron, 1986).

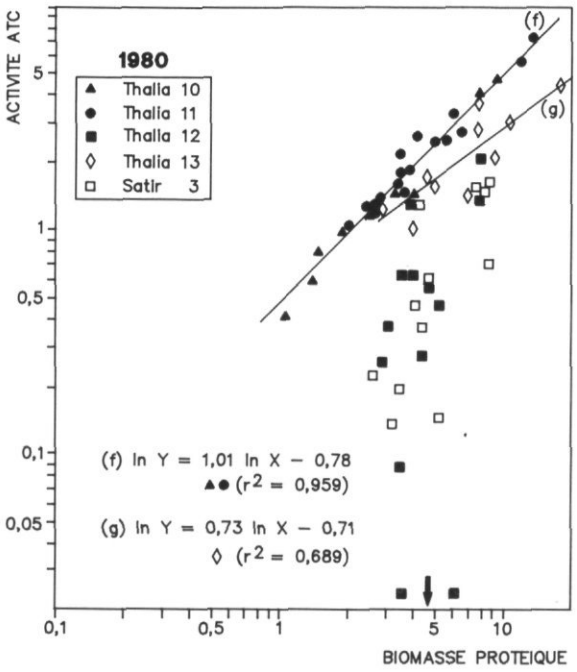

Figure 10. Variations en coordonnées ln-ln (népérien) de l'activité de l'ATC (exprimée en u ATC) en fonction de la biomasse protéique (exprimée en mg) au cours de la période printanière de 1980. Les deux variables sont rapportées au m^3 d'eau échantillonnée. Les calculs des paramètres des droites de régression et des coefficients de corrélation ont été effectués à l'aide des couples de valeurs représentés par les symboles figurant respectivement auprès de chacune des équations.

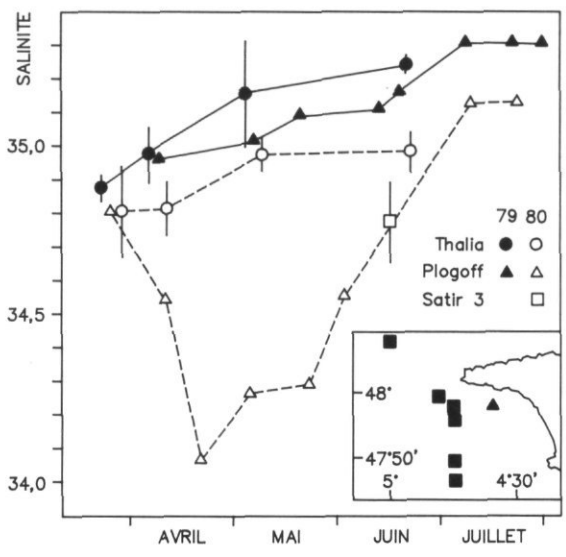

Figure 9. Evolutions saisonnières de la salinité en 1979 et 1980: valeurs moyennes et écarts-types pour l'ensemble des stations; comparaison avec des valeurs (moyenne de 12 mesures pour Satir 3; 11-18/6/80) enregistrées durant les mêmes périodes au niveau de la pointe sud de la Bretagne (positions des stations précisées en cartouche; situation par rapport à la zone d'étude indiquée sur la figure 1).

Conclusion

La mise en évidence de relations allométriques entre l'activité de l'ATC et la biomasse protéique d'un peuplement mésozooplanctonique est un résultat capital au niveau de perception adopté. A ce propos, la sagesse impose de rappeler une assertion de Calder (1985): "One must bear in mind that allometric equations are only correlations. As such they explain nothing, but they do provide the clearest patterns which we seek to explain, hopefully a glass-bottomed bridge from the familiar to that obscured by the diversity in the oceans". Ignorant la diversité du peuplement, la mesure de l'activité de l'ATC tire profit de l'unicité d'un processus métabolique élémentaire commun à l'ensemble des composants de ce peuplement pour en estimer le taux de fonctionnement global (Bergeron, 1983).

Le cas présenté par la campagne Suroît I est particulièrement révélateur. La simple considération des variabilités respectives de la température et de la salinité permet d'identifier clairement trois principaux types de masses d'eau, auxquelles correspondent des teneurs moyennes en chlorophylle a caractéristiques. La répartition coïncidente en trois groupes des valeurs de l'activité de l'ATC mesurées simultanément procède vraisemblablement d'un déterminisme systémique: leur variabilité est étroitement associée à la structure de l'environnement hydrobiologique; elles ne sont directement corrélées ni à la température du milieu, ni aux teneurs en chlorophylle a; elles sont plutôt, par essence, l'expression de l'intégration biologique par le système mésozooplanctonique des influences complexes, encore mal connues du point de vue des effets synergiques ou antagonistes, des multiples facteurs abiotiques et biotiques de l'environnement.

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Claude Leroy, aimable détenteur pour le laboratoire OCEHAL des données de Météo France, Denis Binet (Antenne ORSTOM au Centre de Nantes de l'IFREMER), pour de judicieux commentaires sur le manuscrit, et Louis Giboire, qui s'est employé au dessin des figures, sont bien vivement remerciés.

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Thalassiothrix longissima, a possible oceanic indicator species in the North Sea

P. C. Reid, S. C. Surey-Gent, H. G. Hunt, and A. E. Durrant

Reid, P. C., Surey-Gent, S. C., Hunt, H. G., and Durrant, A. E. 1992. *Thalassiothrix longissima*, a possible oceanic indicator species in the North Sea. – ICES mar. Sci. Symp., 195: 268–277.

Thalassiothrix longissima Cleve et Grunow is primarily an oceanic diatom, but also occurs at times in the North Sea. Its occurrence at 10 m in the eastern Atlantic and North Sea has been monitored by the Continuous Plankton Recorder (CPR) survey at approximately monthly intervals from January 1948 to December 1990. It was common in the North Sea in the period 1952 to 1958 and again in 1967 and 1989; in other years it was not found, or occurred in low numbers. This interannual variability shows a relationship to changing patterns of salinity in the North Sea and appears to reflect the passage of oceanic water into the North Sea in the previous autumn and winter, though presence in the North Sea is also dependent on the size of populations in the North Atlantic. Outbreaks of Paralytic Shellfish Poisoning on the eastern coast of Britain show peaks which may be related to variability in oceanic influx or to the climatic factors contributing to the growth of *T. longissima* in the North Sea. A hypothesis is proposed to explain the above phenomena, where influx of oceanic water is dependent on acceleration of North Sea circulation by north/south movement of the tracks of dominant weather systems.

P. C. Reid, S. C. Surey-Gent, H. G. Hunt, and A. E. Durrant: Natural Environment Research Council, Plymouth Marine Laboratory, Citadel Hill, Plymouth PL1 2PB, UK.

Introduction

The concept that some species or groups of plankton organisms were indigenous to specific "watermasses" and could therefore be used to indicate water movements was first proposed by Cleve (1897) and Aurevillius (1898). Russell (1935a, b, 1936) developed these ideas following his discovery that the chaetognathes *Sagitta elegans* Verill and *Sagitta setosa* J. Muller characterized "western" and "channel" water at the entrance of the English Channel. Russell (1939) applied the concept to the North Sea and documented characteristic indicator plankton from different watermasses, including forms which came in from the Atlantic. Fraser, in a series of papers (e.g. 1955, 1969) was able to show that mixed water passing through the Orkney/Shetland passage contained oceanic plankton which penetrated into the North Sea four to eight weeks earlier than the flow around Shetland. In all the above studies the indicator species used were relatively large zooplankton. Although Russell (1935b, 1936) suggested that larger species which are easily seen without the use of microscopes should be used as indicators, because they can be identified quickly and reproduce more slowly, smaller species have been used successfully.

The spread of a diatom *Biddulphia sinensis*, which appears to have been introduced into the North Sea in about 1903, was charted by Ostenfeld (1908, 1909, 1913); he used it to indicate water movements and in some instances to calculate current speeds. Ostenfeld (1913) also drew attention to the diatom *Thalassiothrix longissima* as a potentially useful indicator of currents, because it is comparatively large and easily recognized. *T. longissima* is an oceanic diatom which has its main centre of distribution and abundance in the Labrador and Irminger current systems of the northwestern Atlantic (Fig. 1). It also occurs in the northeastern Atlantic, often in considerable numbers, and only in certain years extends into the North Sea and Skagerrak, as noted by Ostenfeld (1913). This paper examines these patterns of occurrence, giving special attention to the possibility that years of abundance may indicate an increased influx of oceanic water into the North Sea.

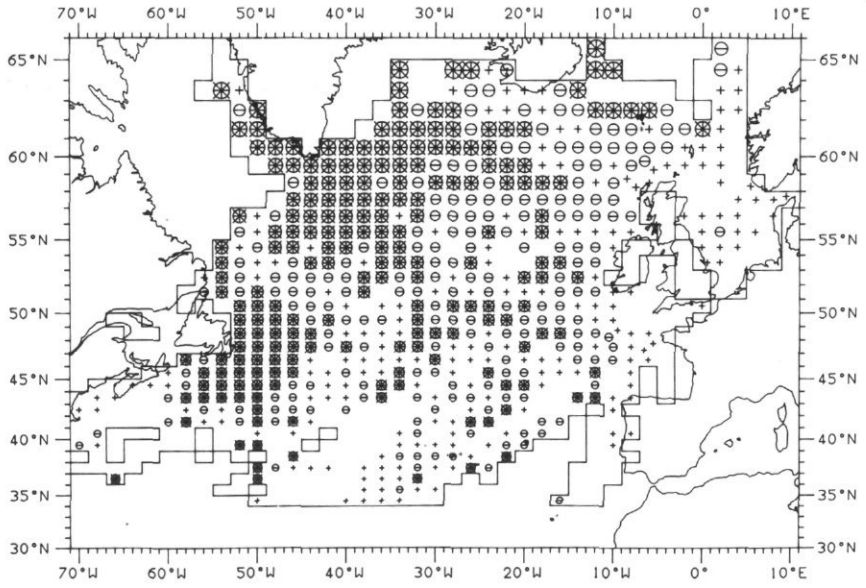

Figure 1. Mean numbers per subsample of *T. longissima* in the North Atlantic and North Sea (1958–1990) averaged for standard rectangles of 2° latitude × 1° longitude. (Symbol intervals + 0.017, 0.200, 0.577.)

Methods

Continuous Plankton Recorders, CPRs (Colebrook *et al.*, 1991) have been used in a routine synoptic survey of the plankton at 10 m depth in the northeastern Atlantic and North Sea at approximately monthly intervals since January 1948. The plankton is sampled on a continuously moving band of silk, of mesh size 270 μm, which is wound on to a storage spool and preserved in formalin. The silk is cut into samples representing 10 nautical miles of tow and the plankton identified and counted on alternate samples. Most of the phytoplankton passes through this coarse mesh; *T. longissima* is one of the exceptions, its long acicular frustules and habit of combining in mats makes it one of the phytoplankton species best sampled by the CPR.

In 1958 a new method of phytoplankton analysis was introduced (Colebrook, 1960) to replace the method used from 1948 to 1957 (Rae, 1952); briefly, 20 microscope fields of 0.3 mm diameter are examined in two diagonals across the filtering silk of each sample; these 20 fields are referred to as a subsample. Abundance estimates are worked up as numbers per subsample which is 1:7551 of a sample silk; taken as 0.0001 by Robinson and Hiby (1980) who give a statistical evaluation of the method. All samples taken over the period 1948 to 1957 in the eastern Atlantic and North Sea (2,401) have been reanalysed for *T. longissima* using the same standard analysis technique to give a continuous time series from 1948 to 1990. Monthly means of all samples were calculated for a set of rectangles (2° longitude by 1° latitude). These were in turn averaged in large standard areas (Fig. 2) and subsequently grouped into eastern Atlantic (areas C5, C4, B5, B4) and North Sea (areas C2, C1, D2, D1) regions.

For comparison with plankton abundance, time series of temperature and salinity have been processed for these same areas with data extracted from the *Bulletin Hydrographique* for the period 1948 to 1957 and for 1958 to 1974 from a series of ICES bulletins of "Monthly means of sea surface temperature and salinities". Charts of mean monthly sea-surface salinity published by ICES for the period June 1950 to December 1958 have been used to measure the surface area covered by water greater than 35 psu, extended to 1973 using coarser estimates derived from the bulletins. Published information is not available after 1974, so it has not been possible to extend these time series to 1990.

Results

The annual fluctuations in abundance for the period 1948 to 1979 of *T. longissima* averaged for the eastern Atlantic and North Sea are given in Figure 3. It was common in the North Sea in 1952 to 1958, occurred in moderate numbers in 1967 and again in 1989, and was not found, or occurred only in very low numbers, in other years. In the Atlantic, *T. longissima* occurred in every year, showing a downward trend in abundance from 1948 to 1980 followed by a moderate increase to 1990. It was a common species in the Atlantic in many of the years when it was not found in the North Sea, for example 1948, 1949, 1960, 1961, 1966, 1968, and 1972.

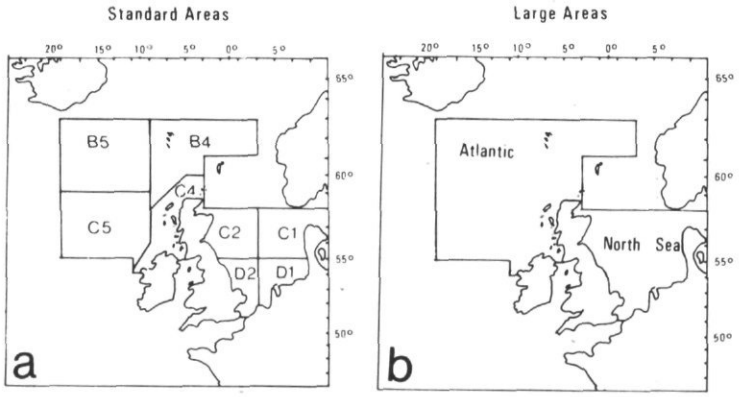

Figure 2. a – The eight standard areas which were used in the data processing of counts of *T. longissima* and averaged to give mean results for b – the two larger areas, the Atlantic and the North Sea.

Figure 4 displays the mean cycle of seasonal abundance for *T. longissima* since 1948 in the two regions; in the Atlantic there tend to be two annual peaks of abundance in the spring and autumn, while in the North Sea the greatest numbers were found in the period from winter to spring with a peak in March. This seasonal pattern was confirmed by a cross spectral analysis on the data for the period 1950 to 1958, when *T. longissima* was sufficiently abundant in the North Sea and North Atlantic for time series of monthly values to be constructed. The analysis shows North Sea populations leading the North Atlantic by one to two months.

There were marked differences in the seasonal patterns of abundance between individual years, with a tendency for persistence of a particular pattern for short series of years; (Fig. 5). In the northeastern Atlantic region, high numbers of *T. longissima* occurred in both

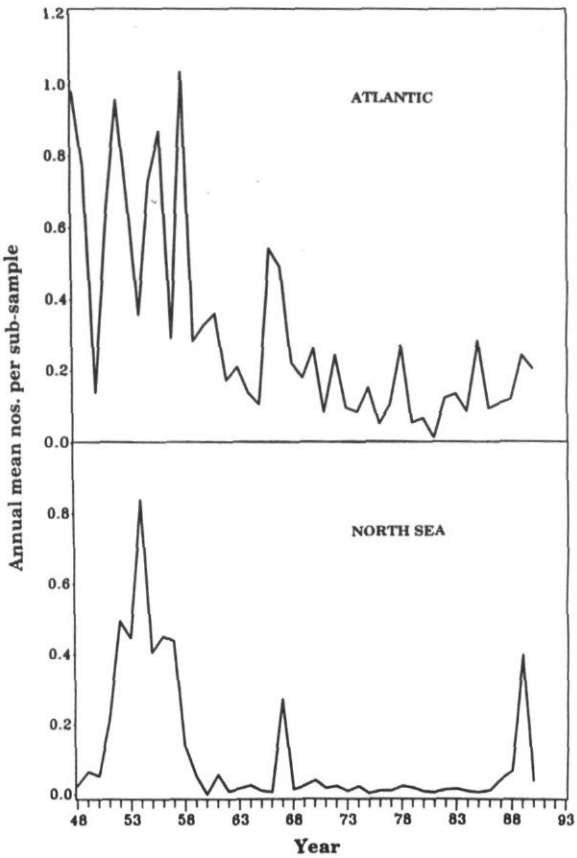

Figure 3. Annual mean abundance of *T. longissima* in the eastern North Atlantic and North Sea areas for the period 1948 to 1990.

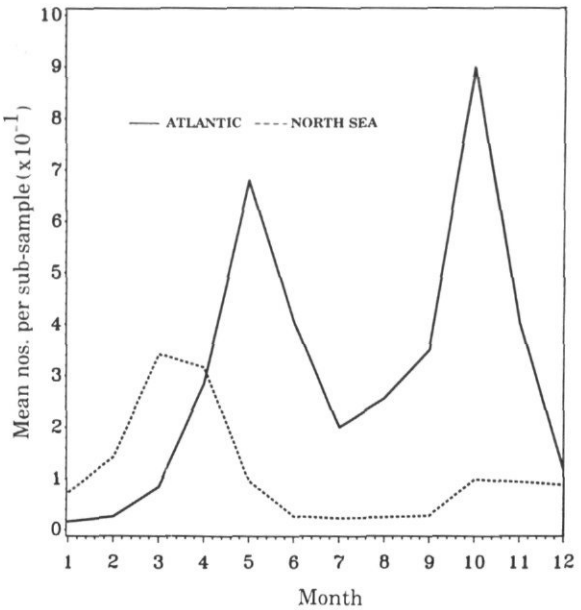

Figure 4. The mean monthly cycle of *T. longissima* (1948–1990) averaged for the eastern Atlantic and North Sea.

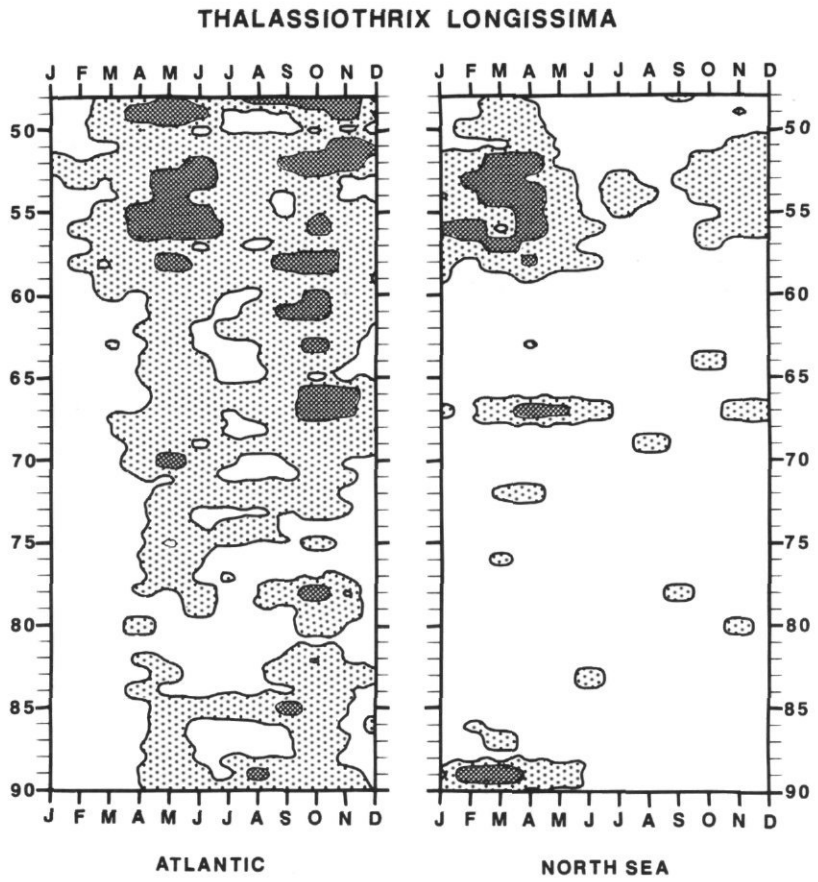

Figure 5. Contour diagrams of the monthly mean abundance of *T. longissima* in the eastern North Atlantic and North Sea with months across and years down. Contour levels: >0, 0.1, 1.0 (from palest to darkest areas).

the spring and autumn from 1948 to 1958, with the exception of 1950, followed by four years (1960 to 1963) with low numbers in the spring but higher numbers in the autumn. Subsequent years were largely characterized by reduced numbers in both the spring and autumn, with the exception of an intensive bloom in the late autumn of 1966 and 1967. In the North Sea, numbers were low in the first few years, 1948 to 1951. This was followed by a period from December 1951 to May 1958 with high numbers in the late winter and spring. *T. longissima* almost disappeared from the North Sea in the subsequent period until 1967 when it was abundant in the early spring to be followed again by low numbers until a further increase in the spring of 1989.

Examination of the monthly means for the two regions plotted as a time series (Fig. 6) shows that in the period 1948–1957, when *T. longissima* was abundant in the Atlantic, eight out of ten late autumn increases in the Atlantic were followed by a spring increase >0.5 in the North Sea. This pattern was broken in 1958, the year with the highest recorded autumn Atlantic population when a spring bloom failed to materialize in the North Sea. The annual population size was generally lower in the Atlantic in subsequent years (Fig. 3) during which period 11 years with Atlantic autumn peaks >0.5 were not followed by a bloom in the North Sea and 19 years showed low late autumn Atlantic blooms again with no spring expression in the North Sea. An Atlantic late autumn population peak in 1966 was followed by a spring increase in the North Sea in 1967, but the almost equally large autumn population in 1967 failed to produce a North Sea bloom as in 1958. The year 1989 is exceptional in that a sizeable spring increase in the North Sea did not appear to follow an Atlantic autumn peak in the previous year. A significant correlation between the time series for the Atlantic and the North Sea (0.52 $n = 43$ $p < 0.001$) shows that populations in both areas are following a similar trend. This is confirmed by a cross-spectral analysis of the period 1952 to 1958, which shows that the trend in the Atlantic precedes that in the North Sea by several months.

In years of abundance the late autumn/winter increase of *T. longissima* in the North Sea was typically followed by a sharp increase to high numbers in the following spring. This spring increase cannot have originated from the Atlantic population at this time of year and must

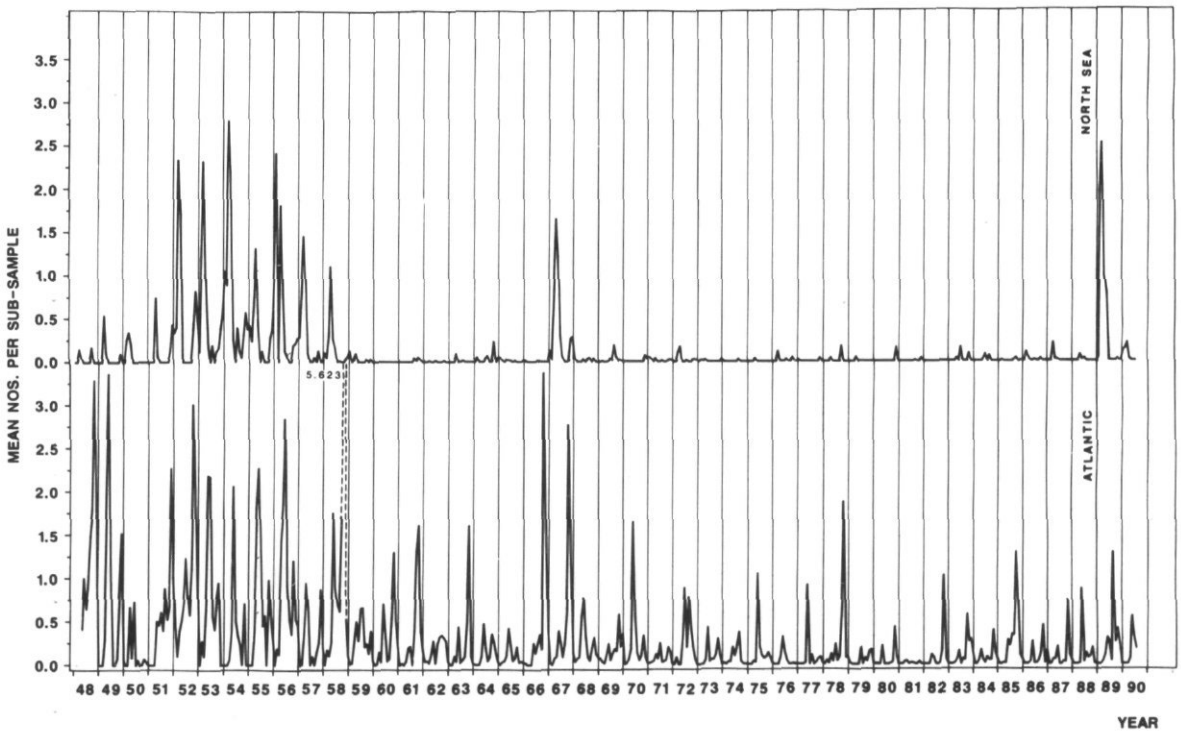

Figure 6. Monthly means of *T. longissima* in the eastern North Atlantic and North Sea from January 1948 to December 1989 plotted as a time series.

represent rapid growth of the species in the North Sea, possibly seeded from the autumn influx, a view that is supported by the cross-spectral analysis. The alternative possibility of a development of the North Sea spring population from resting stages is unlikely because of the irregular nature of the occurrence of a species which is also not known to form resting stages.

A major outbreak of Paralytic Shellfish Poisoning (PSP) which led to the hospitalization of 78 people was recorded on the northeast coast of England in 1968 (Ayres, 1975). Monitored since then using a mouse bioassay, PSP has been found to be present every year, but did not produce high mouse unit titres again until 1990 (Fig. 7). This interval corresponds with a one-year lag to the interannual variability of *T. longissima* for this period and the causes of the two phenomena may be related.

Discussion

The seasonal cycle of *T. longissima* when it occurs in the North Sea is unusual in that it is predominant in the winter and early spring months of the year, when most other plankton species are found only in small numbers. It is a large diatom and may require turbid conditions to remain in the photic zone in those months of the year when the water of the North Sea is completely mixed. As

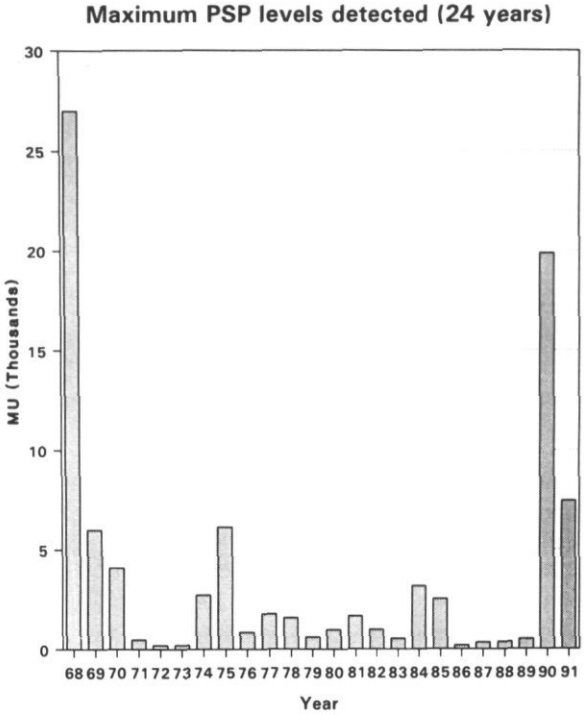

Figure 7. The maximum level of Paralytic Shellfish Poisoning detected on the northeast coast of England in the period 1968 to 1991. Toxicity expressed in mouse units (MU) per 100 g shellfish tissue. Figure provided courtesy of MAFF.

soon as stratification develops in the northern North Sea (May), *T. longissima* disappears from the plankton at 10 m, and the autumnal occurrences do not commence until the breakdown of stratification in October/November.

The unusual seasonal cycle in the North Sea, and especially the peak which characterizes the spring bloom, may be related to the seasonal cycle of salinity. The average maximum extent of 35 psu salinity water in the North Sea coincides within one month with the spring maximum of *T. longissima*, and the rapid decline in numbers of this diatom in the late spring follows the western spread of meltwater from the Norwegian coast and the Baltic after the spring thaw.

The occurrence of *T. longissima* in rectangles of 2° longitude × 1° latitude is plotted in Figure 8 against sea-surface temperatures and salinity for these same areas using 1957, when the species was abundant in the North Sea, as a representative year. The oceanic and North Sea populations are clearly distinguished, reflecting the different seasonal combinations of temperature and salinity. In 1957 the species was found in waters with a wide range of temperatures and salinities, but it is not known if it was actively growing throughout the whole of this range. Ostenfeld (1913) gave optimum conditions for growth of *T. longissima* as being in the range 5.4–11.8°C and 25.33–35.96 psu with means of 8.7°C and 35.31 psu. These compare with means of 9.6°C and 34.48 psu determined by Cleve (1900). Our comparisons with salinity from 1957, which are based on coarse data averaged for large sea areas, show that all Atlantic records and the greater majority of records in the North Sea were found at salinities greater than Cleve's (1900) mean of 34.48 psu, implying that the diatom grows best in higher salinities. Ostenfeld's (1913) suggestion that a growth optimum of approximately 9°C may be the cause of its disappearance in the summer in the North Sea seems unlikely, as it was found in the Atlantic at temperatures up to 14°C and up to 18.2°C by Cleve (1902).

The monthly extent of 35 psu water in the North Sea measured from charts of surface salinity from 1950 to 1973 is given in Figure 9. A seasonal pattern is repeated nearly every year, with a maximum areal coverage in February to April, a sharp decline in the summer months and a gradual increase from October. High salinity water (>35 psu) varied greatly in extent; it was exceptionally widespread in four of the years (1954 to 1957) when *T. longissima* was abundant in the North Sea, with the highest numbers in the North Sea in the year with the widest spread of 35 psu water, 1954. Water with an exceptionally high salinity had also penetrated the northern North Sea in early 1990 (Heath *et al.*, 1991) coinciding with the re-appearance of *T. longissima* and the occurrence of other oceanic plankton (Lindley *et al.*, 1991). It is clear, however, from a comparison of Figures 6 and 9, that conditions of salinity suitable for the growth of the diatom are likely to have occurred in the spring of

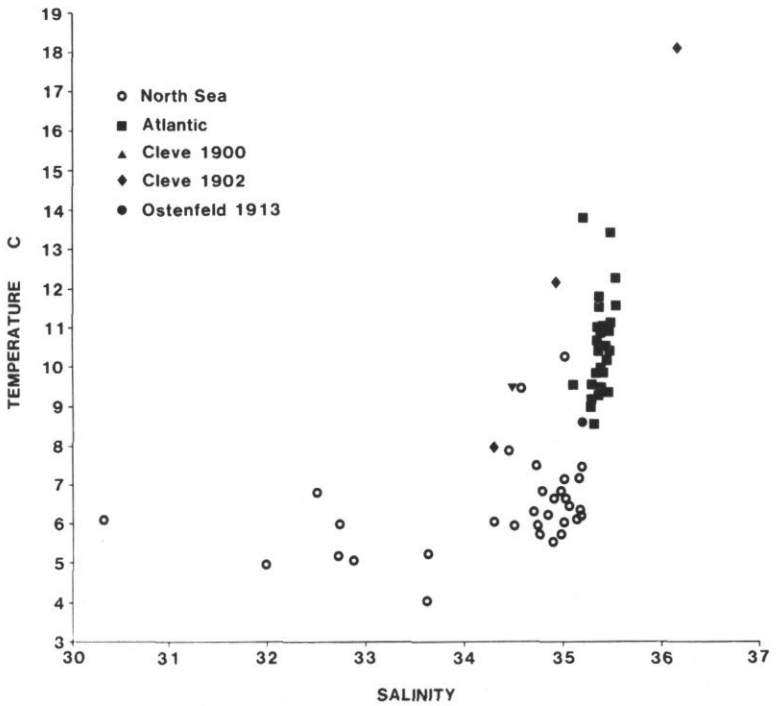

Figure 8. Occurrence of *T. longissima* in 1957 by 2° longitude × 1° latitude rectangles within the North Atlantic and North Sea areas in relation to monthly mean temperatures and salinities for the same rectangles with additional observations on the temperature/salinity relations of this species from Cleve and Ostenfeld.

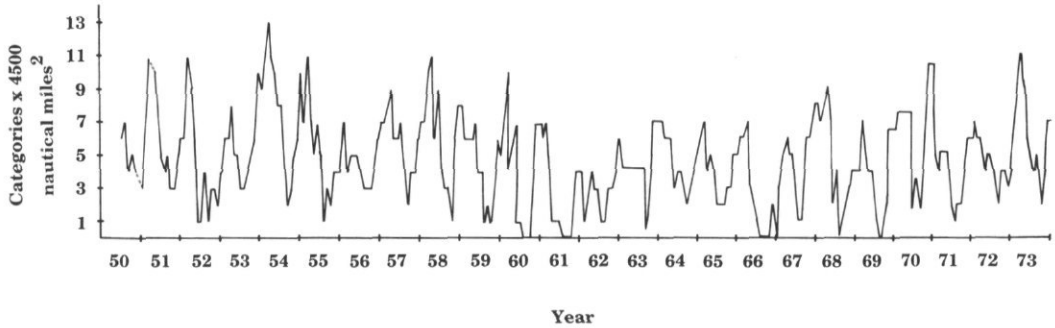

Figure 9. Areal extent in square nautical miles of 35 psu water for the North Sea south of 58°N measured from charts (1950–1958) and bulletins of rectangle means (1959–1973) of sea-surface salinity published by ICES.

nearly every year. While the coincidence of high salinity water in the North Sea with blooms of *T. longissima* in the spring appears necessary for its growth, it does not explain why it does not grow in other years, and especially 1973 when suitable salinities in the spring almost certainly occurred. Other factors, such as river runoff, rainfall, and evaporation, also affect salinity (Taylor *et al.*, 1983), so the extent of 35 psu water is only partly due to the advection of oceanic water into the North Sea.

The absence of *T. longissima* from CPR samples in the North Sea in 1960 followed a failure of the autumnal bloom in the Atlantic in 1959. Low autumn populations in the Atlantic may, at least, be the partial cause of the failure in the North Sea of *T. longissima* to increase in the spring in many years, e.g. 1949/1950, 1950/1951, 1962/1963, 1964/1965, 1965/1966, and all years from 1968 to 1979, years which include 1951 and 1973 when extensive areas of 35 psu water were found in the North Sea. In 1959 *T. longissima* showed a marked decline in numbers in both the eastern Atlantic and North Sea after seven years of abundance. A contributory cause of this failure may have been the changing meteorological conditions that prevailed in this year (Namias, 1964), when exceptionally high sea-surface temperatures occurred in the North Sea and gave rise to stable conditions and strong stratification much earlier in the year than usual (Filiarski *et al.*, 1959). This early stratification was associated with the occurrence of unusually large numbers of *Ceratium* spp. in the spring months of the year; especially in area C1 (see Reid, 1978a) such stable conditions are likely to be unsuitable for *T. longissima* growth.

The downward trend from 1948 to the early 1980s and subsequent recovery in the abundance of *T. longissima* (Fig. 3) in the eastern Atlantic is similar to annual fluctuations in zooplankton in this same area (Colebrook *et al.*, 1991; Colebrook, 1991). The pattern confirms Colebrook's contention that phytoplankton follows a similar trend, by extending a phytoplankton data series back to the same start date as the zooplankton. Correlations with atmospheric climatic variables by Colebrook and with the strength of the Gulf Stream by Taylor *et al.* (in press) suggest that the trend is mediated by climatic variability, though the mechanisms of any relationship are not yet understood.

Reid (1978b) showed that annual mean surface salinities in the entrance to the Skagerrak and in the eastern North Sea were inversely related to the depth of the 35 psu isohaline in the central Skagerrak (Fig. 10). It was suggested that low salinity outflows were compensated by higher salinity inflows along the Norwegian Rinne. Furnes (1992) has eloquently confirmed this suggestion with his modelling calculations in the JONSDAP Utsira section after March/April 1976. Years when *T. longissima* was most abundant in the North Sea, 1952 to 1959 and 1967 (Fig. 3), coincided in a negative sense with a lag of one year to the surface salinity at the mouth of the Skagerrak and in a positive sense with increasing proximity of 35 psu water to the surface in the Skagerrak (Fig. 10). After 1968 this pattern breaks down, a response which may have been due to low Atlantic populations which were not sufficient to seed the North Sea. An inverse relationship (Fig. 11) has also been demonstrated between the salinity in the western and eastern central North Sea (Reid and Budd, 1979; Reid, 1990); over the period 1948 to 1974 the correlation is $r = -0.56$ $n = 27$ and increases to -0.64 $n = 26$ (both $p < 0.001$) if area C2 is lagged by one year in front of C1. Monthly variability in the salinity of areas C1 and C2 is plotted in Figure 12 for the period 1948 to 1974.

The influx of Atlantic water and the circulation of the North Sea have been shown to be closely dependent on the variability of weather and in particular wind stress and direction, with greater inflow of oceanic water following southwest to northwest winds (Furnes, 1992; Svendsen and Magnusson, 1992). Surface salinities in the entrance to the Skagerrak and in the eastern North Sea are controlled by the strength of the Baltic outflow of low salinity water, Norwegian coastal runoff and the spread of Norwegian coastal water by offshore winds.

We propose a hypothesis which attempts to explain the interannual variability of salinity and *T. longissima* described above as a systematic response to regional climatic changes. The residual circulation of the North

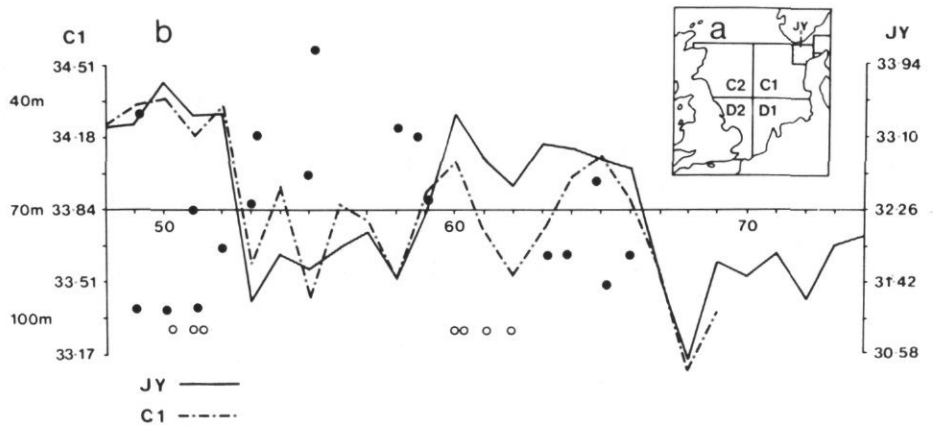

Figure 10. a – Rectangle JY and North Sea standard areas. b – Standardized plots of annual mean surface salinity based on a mean of 27 years for rectangle JY and a mean of 22 years for standard area C1, with superimposed the depth of the December–February 35 psu isohaline in the central Skagerrak, replotted from Figure 11 in Svansson (1975) and updated to 1966 from hydrographic reports published in *Annales Biologiques*. Open circles represent depths greater than 100 m.

Sea is anticlockwise and may be reinforced by dominant westerly winds when depressions have a northerly track as in 1989 and 1990. Influx of Atlantic water down the west side of the North sea is compensated by outflow of Baltic and coastal water along the Norwegian coast. These depressions also cause higher rainfall over Scandinavia and increased runoff, which will contribute an additional density-driven component to the Baltic outflow and Norwegian coastal current. River runoff statistics provided by the Norwegian Water Resources and Energy Administration for the River Vossa confirm exceptionally high outflow during 1989/1990. It is proposed that the North Sea acts like a venturi tube with water flowing out in the Norwegian current pulling in more oceanic water down the west side of the North Sea with a slightly lagged compensating inflow forming later as a subsurface current in the Norwegian Trench. In years when depression tracks are further south, the general circulation and oceanic influx to the North Sea is likely to be reduced. *T. longissima* as an oceanic indicator supports this thesis until 1967. Its subsequent absence until 1989 may be a reflection of lower populations in the Atlantic, implying that a sizeable influx of *T. longissima* into the North Sea is required for a successful spring bloom to be seeded.

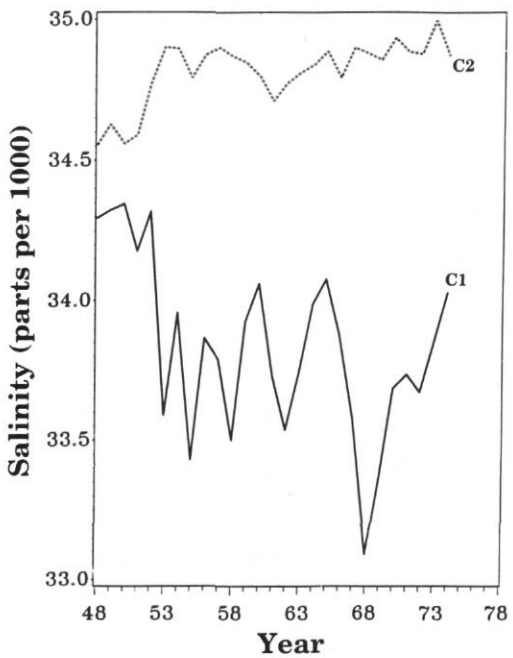

Figure 11. Annual mean sea-surface salinity for the period 1948 to 1974 averaged for areas C2 in the west central North Sea and C1 in the east central North Sea (see Fig. 2).

Outbreaks of Paralytic Shellfish Poisoning on the English east coast occurred in years when the maximum salinities measured in the North Sea since 1920 (August 1968 and January 1990) by the Marine Laboratory, Aberdeen were made (Heath *et al.*, 1991) and followed one year after peaks in numbers of *T. longissima*. These outbreaks may be a response to preconditioning by Atlantic water in the western North Sea and/or to climatic forcing associated with the oceanic inflows favouring excystment of resting stages and growth of the dinoflagellate *Alexandrium tamarense* (Lebour) Balech.

Conclusions

Although the evidence is by no means unambiguous, it seems likely that the presence of *T. longissima* does indicate the passage of oceanic water into the North Sea in the autumn months of the year, though its abundance appears to be highly dependent on the size of populations in the Atlantic and may not indicate variability in the volume or rate of transport of oceanic water into the

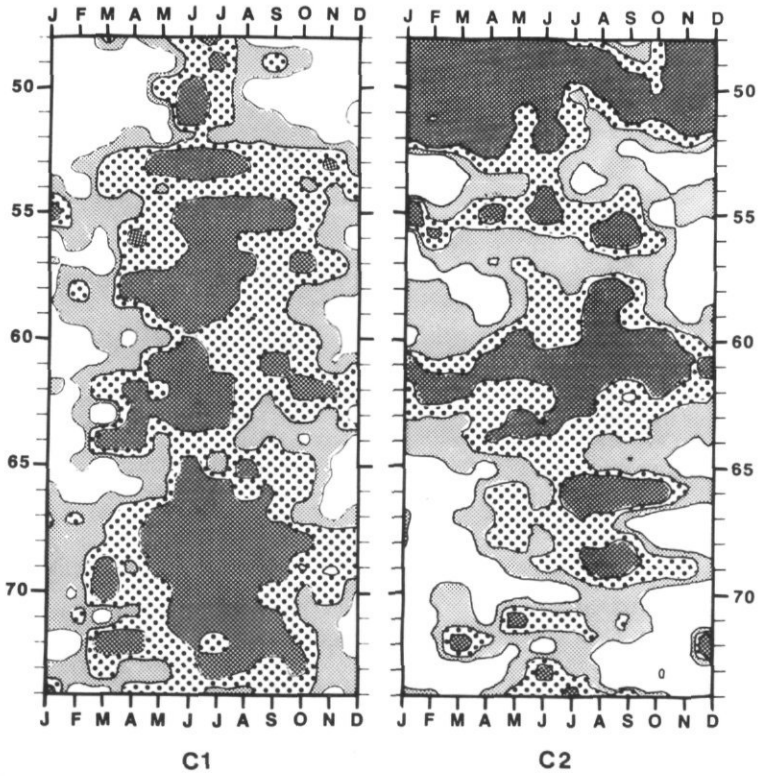

Figure 12. Contour diagrams of monthly mean salinity in areas C2 (west central North Sea), contour levels 34.91, 34.84, 34.76, and C1 (east central North Sea), contour levels 34.31, 33.99, 33.46 (from darkest to palest areas).

North Sea. *T. longissima* does not appear to be able to survive as a permanent member of the North Sea plankton without regular reintroductions.

Changes in *T. longissima* abundance over time appear related to the interannual variability of salinity in different areas of the North Sea. It is proposed that an acceleration/deceleration of the anticlockwise circulation of the North Sea by north/south movement of dominant weather tracks causes increases/decreases of oceanic influx reflected in salinity. These patterns may be behind the erratic occurrence of outbreaks of PSP on the northeast UK coast and are also likely to have an impact on the planktonic productivity of the North Sea and the occurrence of algal blooms.

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The Barents Sea ecosystem in the 1980s: ocean climate, plankton, and capelin growth

Hein Rune Skjoldal, Harald Gjøsæter, and Harald Loeng

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The 1980s was characterized by large ecological fluctuations in the Barents Sea ecosystem. A long period of cold ocean climate from 1977 to 1981 ended with a major inflow event of Atlantic water during 1982 and early 1983. A second major inflow event occurred in 1989 and 1990. The first major inflow event ended in late winter and left the southwestern Barents Sea filled with water containing low abundance of the dominant copepod *Calanus finmarchicus*. This reflects the large-scale seasonal vertical migration whereby *C. finmarchicus* overwinters in the adjoining Norwegian Sea deeper than the depth at the entrance to the Barents Sea. This mechanism caused a minimum in zooplankton biomass in the central Barents Sea in 1984, which again resulted in a poor individual growth rate of capelin and contributed to the sharp decline of the capelin stock from 1984 to 1986. Coinciding with the decline of the capelin stock, there was an increase in zooplankton biomass up to a maximum around 1987. Released predation pressure from capelin was probably a major reason for this increase. High abundance of zooplankton, particularly of larger forms such as krill and amphipods, is considered to be a major factor which caused very high growth rate and rapid recovery of the capelin stock up to 1990. The 1980s have provided us with examples of the dynamics of the Barents Sea ecosystem and the complex and intricate relationships which exist between ocean climate variability as a primary driving force, zooplankton stocks, and growth rate and population dynamics of capelin. Only through insight into the basic mechanisms can such relationships be properly described and used for future predictive purposes.

Hein Rune Skjoldal, Harald Gjøsæter, and Harald Loeng: Institute of Marine Research, PO Box 1870, Nordnes, N–5024 Bergen, Norway.

Introduction

The 1980s was a period of dramatic fluctuations in the Barents Sea ecosystem. The most notable event was the collapse of the capelin stock to a very low level in 1986 and its subsequent rapid recovery back to a high level by 1990 (Hamre, 1988, 1991, 1992; Skjoldal and Rey, 1989). The low abundance of capelin, which is a key intermediate link in the food web, had a drastic effect on higher trophic levels. Among the effects were massive invasions of harp seals along the Norwegian coast, high mortality of seabirds, and poor individual growth of cod (Mehl, 1989, 1991; Hamre, 1991, 1992; Skjoldal, 1990; Blindheim and Skjoldal, 1992).

The 1980s was also a period of intensified research in the Barents Sea ecosystem. There has been extensive data collection on capelin and other fish stocks through national and joint Soviet–Norwegian surveys. Plankton data have been collected by the Institute of Marine Research with the purpose of describing feeding conditions of capelin since 1979. This research was intensified and broadened with the Norwegian Research Programme on Marine Arctic Ecology (PRO MARE) in 1984–1989. Oceanographic data are collected regularly on fixed sections and are part of a long time series that for the Kola section goes back to the beginning of this century.

The extensive research and monitoring activities have left us in a good position to examine and analyse in retrospect the ecological events that took place during the 1980s. The climatic events have been reviewed by Loeng *et al.* (1992). Hamre (1991, 1992) has reviewed the changes in the capelin stock size and considered mechanisms responsible for these changes involving interactions between the capelin, herring and cod stocks. Here we describe and examine the fluctuations in

ocean climate and zooplankton biomass and show how these factors were related to variations in growth rate and stock size of capelin.

Material and methods

The Institute of Marine Research has maintained observations in several standard sections in the Barents Sea in order to map climatic signals in the current system. Since 1977 observations have been carried out six times a year at the Fugløya–Bjørnøya section and four times a year at the Vardø-N section (Fig. 1). The longest time series of temperature in the Barents Sea is the Russian recording in the Kola section (Fig. 1), which started as early as 1900. The data from the Kola section have been provided by PINRO.

Means of temperature and salinity have been calculated for the main Atlantic inflow through the various sections using the method described by Blindheim and Loeng (1981). These means are prepared for the depth interval of 50–200 m along the parts of the sections which are marked in Figure 1. In the Kola section, means from 0 to 200 m are used. By using means over part of sections, small-scale spatial variability is filtered out.

Zooplankton biomass data have been collected by the Institute of Marine Research since 1979. Data for the period 1979–1984 were reported by Skjoldal *et al.* (1987), and details on methods are given there. Up to 1982 the biomass was recorded as zooplankton volume and these data have been converted to dry weight and ash-free dry weight using standard conversion factors. The samples have been collected with vertical net hauls (Juday or WP-2 nets, 180 μm mesh size in most cases), with additional samples collected with MOCNESS multiple net sampler (333 μm mesh size) since 1984.

There has been a shift in time of sampling from mainly the summer season up to 1984 to more samples from the autumn season from 1986 onwards. This reflects an emphasis on processes associated with the receding ice edge and summer feeding of capelin in the early period and data collection during the autumn multispecies surveys from 1986. The large data set on zooplankton from the Barents Sea during the 1980s is being analysed and will be reported more extensively elsewhere (H. R. Skjoldal, A. Hassel, and L. Omli, unpublished results).

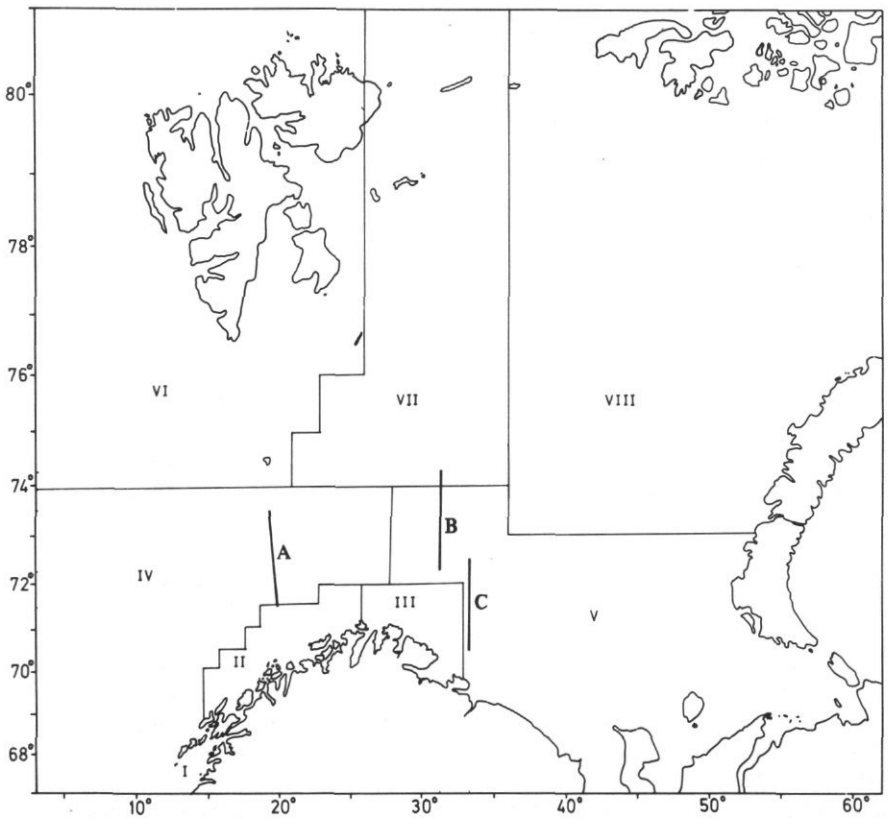

Figure 1. Map of the Barents Sea showing the MULTSPEC subareas used for calculating mean zooplankton biomass and capelin size and growth rate. The positions of three hydrographic sections are also shown. A: Fugløya–Bjørnøya; B: Vardø-N; C: Kola section.

Here we have used calculated mean biomass values for different subareas used in MULTSPEC, a multispecies model being developed for the Barents Sea (Bogstad and Tjelmeland, 1990; Fig. 1). Because most of the research in the first part of the 1980s was conducted in subarea 7 in the central and northern Barents Sea, data from other subregions are available mainly for the latter part of the 1980s.

Observed weight and length of capelin (*Mallotus villosus*) were analysed based on data from the annual joint Soviet/Norwegian capelin surveys in the autumn. Length- and weight-at-age representative for the whole stock were extracted from detailed stock size tables in the survey reports (Institute of Marine Research, Bergen, unpublished reports). Measurements representative for the MULTSPEC subareas were extracted from tables of stock size for each of these subareas, calculated by an alternative program for acoustic stock size estimation (Gjøsæter, 1985). In both cases, the two-year-old age group has been used to construct time series, for three reasons. First, this age group is unaffected by any possible size-selective spawning mortality. Second, the high number of individuals of this age group sampled each year ensures a good database. Third, this age group has experienced only three growth seasons, and its size therefore reflects cumulatively the growth conditions in the current and the two previous seasons.

Data of this type illustrate accumulated growth over more than one year. This resolution of growth data is insufficient for many purposes, and growth during the current growth season is a better variate to use to reveal interannual variability. The growth in the current season may also to some degree depend on the previous growth history of the fish, and it is in addition difficult to estimate. We have used two different approaches to derive such estimates. One method has been to subtract the mean length and weight of one-year-old individuals observed by the end of the former growth season from the values for two-year-old individuals observed at the end of the current growth season. If size-dependent differences in mortality during the season are small, this method may give reliable estimates of growth for the total stock. The method has clear limitations for revealing growth differences related to geographical distribution because of the extensive migration of capelin.

The second approach is based on a linear relationship between otolith growth and capelin growth in length (Gjøsæter, 1986). The width of the outer growth zone in the otolith is therefore related to capelin length increment in the current growth season. If one assumes that an individual caught in a certain geographical area in late autumn has spent most of the growth season in that area, such estimates of fish growth can be related to measured environmental factors in the subareas. Length increments in the current growth season were therefore calculated from otolith growth zone measurements from individuals inhabiting the various subareas.

Results

Ocean climate

In order to compare the temperature conditions in the Barents Sea during the 1980s with earlier periods, a three-year running mean is shown in Figure 2. The period from 1977 to 1982 was the coldest observed in the Barents Sea since the beginning of the century. Figure 3 shows more details in the variability in temperature and salinity of the Atlantic inflow to the Barents Sea during the last 15 years. The cold period lasted to the beginning of 1982, but the temperature had started to increase already during the summer of 1981 in the western part of the Barents Sea. In the eastern part the warming started later, during the spring of 1982 (Loeng *et al.*, 1992). The warm period lasted for three years (1982–1984) and was followed by four years with temperatures just below average (1985–1988). In 1989 there was a rapid temperature increase in the whole southwestern Barents Sea (east to about 35°E). Since then the temperature has been almost 1C° above the average. In the eastern

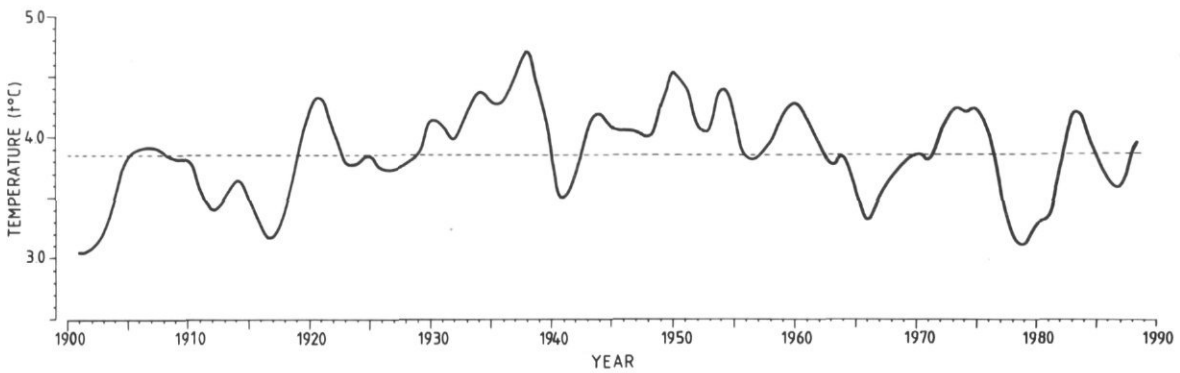

Figure 2. Three years' running average of yearly temperature in the Kola section during the period 1900–1990, based on data from Bochkov (1982) and from PINRO, Murmansk (pers. comm.). From Loeng (1991).

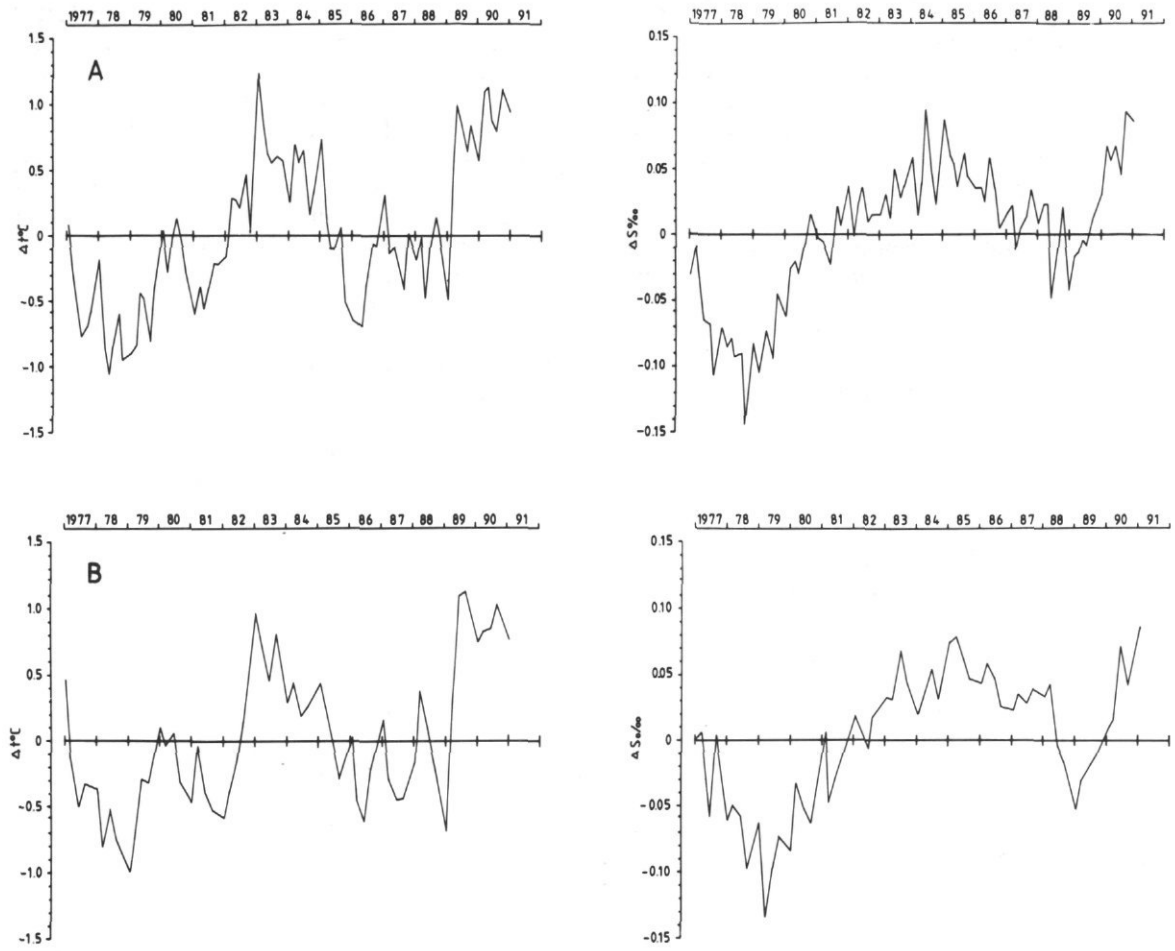

Figure 3. Anomalies of temperature (left) and salinity (right) in the sections Fugløya–Bjørnøya (A) and Vardø-N (B). From Loeng *et al.* (1992).

Barents Sea the temperature increase took place during the winter and spring of 1990.

The salinity reached a minimum in 1978/1979 (Fig. 3). This marked the arrival of the large salinity anomaly that was advected around in the northern North Atlantic during the 1960s and 1970s (Dickson *et al.*, 1988; Blindheim and Skjoldal, 1992). There was a marked increase in salinity up to 1980, followed by a slower increase until maximum was reached in 1984. The salinity decreased to a minimum in 1989, followed by a new increase. The main features in the variability of salinity follow the temperature fluctuations, but Figure 3 also indicates small differences.

Zooplankton biomass

A time series of mean zooplankton biomass obtained with vertical net hauls in subarea 7 (Fig. 1) reveals a minimum in the zooplankton biomass in 1983 and 1984 (Fig. 4). The data up to 1984 have been presented in more detail by Skjoldal *et al.* (1987) and Rey *et al.* (1987). The minimum in 1984 reflected low abundance of the dominant copepod *Calanus finmarchicus,* which showed a 50-fold decrease from peak numerical abundance in 1982 (Skjoldal *et al.*, 1987). After the minimum in 1984 there was an increase in biomass to a maximum in this data set in 1987, followed by a decrease up to 1990.

Size-fractioned biomass data obtained with vertical net hauls (left column) or MOCNESS plankton sampler (right column) from 1985 to 1990 are shown in Figure 5. The highest mean biomass has typically been found in subarea 8 (northeastern Barents Sea) and the lowest in subarea 5 (southeastern Barents Sea, Fig. 1). The mean biomass in subarea 7 has typically been intermediate between those of areas 8 and 5 (Fig. 5). The total biomass obtained with vertical net and MOCNESS sampler show in most cases quite similar levels. Since the number of samples and the geographical coverage in each subarea have varied considerably, detailed in-

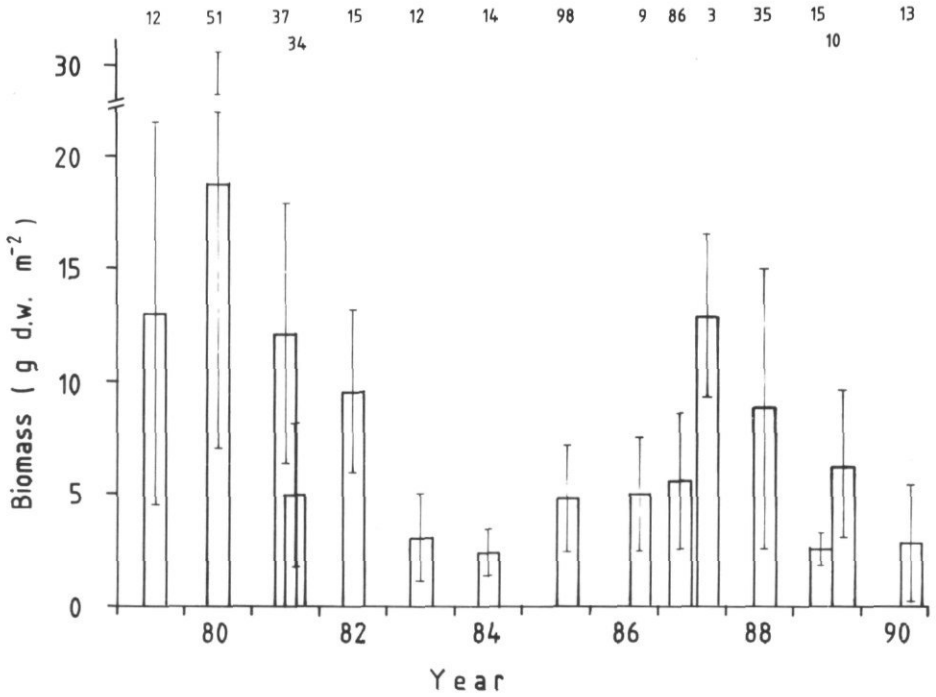

Figure 4. Mean zooplankton biomass for vertical net hauls from stations in the MULTSPEC subarea 7 in the Barents Sea. Vertical lines denote twice the standard deviation, with numbers of stations given at top of the figure. Based on cruises in July 1979, June 1980, June and August 1981, June 1982–1984, August 1985, September/October 1986–1990, and May 1987 and 1989. Based on data from Skjoldal *et al.* (1987) and Skjoldal, Hassel, and Omli (unpubl. results).

terpretation of the results must be made with some caution. In general, the smaller meshed vertical nets give higher biomass in the size fraction <1 mm, whereas MOCNESS collects more of the organisms >1 mm. The size fractions >1 mm were typically dominant in contributing to the total zooplankton biomass.

We have little data available to determine whether the zooplankton minimum in subarea 7 in 1983–1984 also occurred in other subareas. In subarea 5 the biomass showed an increasing trend from low levels in 1987. In subarea 8 the biomass was relatively high from 1985 onwards (Fig. 5).

Capelin growth

Time series of mean length and weight of two-year-old capelin for the total stock show that the growth varied considerably in the period from 1977 to 1991 (Fig. 6a). Length and weight fluctuated in parallel, with minima in 1978, 1984, and 1991. The conditions for growth were better in the period 1980–1983 and were particularly good in the period 1986 to 1990. The weight probably reflects the growth conditions in a more direct way than the length does; it also fluctuates to a larger extent.

Estimated mean growth in length and weight during the current season for the total stock is shown in Fig. 6b. These growth estimates were calculated by subtracting the observed length and weight of one-year-old individuals the previous year from those of 2-year-old individuals in a given year. These data should more accurately reflect the growth of the two-year-olds in the year indicated, since the growth at the 0- and 1-group stages of the same year class is subtracted. The variation in growth in current season is much greater than for mean length and weight (Fig. 6), as was to be expected. The maxima in the periods 1982–1983 and 1986–1990 are more pronounced. Low growth rate, corresponding to weight increments of 5 grams or less, were observed in the years 1977–1981, 1984–1985, and 1991.

Time series of mean length and weight for two-year-old individuals from subareas 5, 7, and 8 are depicted in Figure 7. The pattern of changes in mean length and weight for subarea 7 from 1979 to 1990 is quite similar to that for the total stock (Figs. 6a and 7a). Estimated length increments in the current growth season calculated from otolith growth zones are shown in Figure 8. The growth pattern obtained from otoliths for the total stock revealed the same main features as the pattern derived from length increments from 1 to 2 years of age ($r = 0.78$; Figs. 6b and 8a). However, there were some discrepancies between the two sets of estimates. Thus the otolith method showed higher growth in 1980 and lower growth in 1990 than the observed length increments. The growth pattern from otoliths for subarea 7

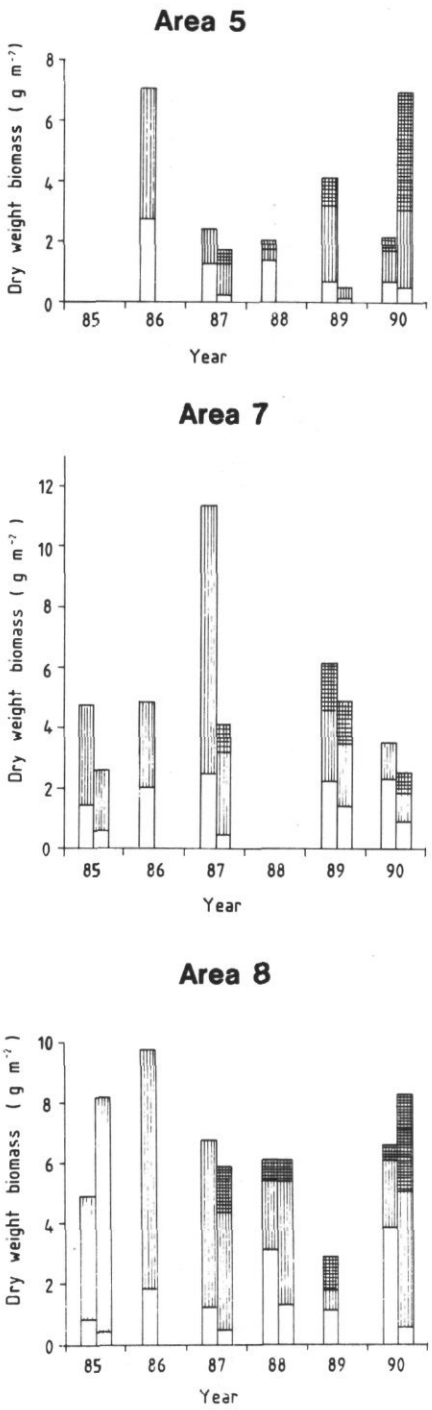

Figure 5. Mean zooplankton biomass for samples obtained with vertical net hauls (left columns) and MOCNESS (right columns) for stations in MULTSPEC subareas 5, 7, and 8. The biomass is given for different size fractions as follows: open columns ≤ 1 mm; vertically hatched columns ≥ 1 mm or 1–2 mm; cross-hatched columns ≥ 2 mm. Samples of the >2 mm fraction were not taken in 1985–1986 and for the vertical net hauls in 1987. From Skjoldal, Hassel, and Omli (unpubl. results).

showed very close correspondence to that for the total stock (Fig. 8a,c). The otolith growth estimates were typically higher for subarea 5 and lower for subarea 8 than the estimates for the total stock, and their pattern of interannual variability also showed some differences (Fig. 8a,b,d).

Discussion

The variability in temperature and salinity is due to fluctuations in the Atlantic inflow to the Barents Sea. The variability in the currents may be explained by external forcing, but may also be a result of processes occurring in the Barents Sea itself (Midttun and Loeng, 1987). One process of importance is the formation and outflow of bottom water from the eastern basin, a process described in detail by Midttun (1985). The outflowing bottom water has to be replaced by inflowing Atlantic water from the west. The outflowing volume may vary considerably from year to year, as will the corresponding inflow. Between 1982 and 1983 and between 1989 and 1990 there were extremely large outflows of bottom water from the eastern Barents Sea which coincided with increased Atlantic inflow (Loeng *et al.*, 1992).

Ådlandsvik (1989) and Ådlandsvik and Loeng (1991) have shown that the variability in the Atlantic inflow is closely related to the wind conditions. The warm periods in the Barents Sea are related to stable low pressure situations in the area, while cold periods are linked to high pressure situations. It is possible that wind-driven inflow during periods of stable low pressure will push the bottom water in the eastern basin out of the Barents Sea and into the Arctic Ocean. Results from a laboratory model support this assumption (McClimans and Nilsen, 1990). The combination of an increased inflow caused by the atmospheric circulation and the associated outflow of bottom water allows for large and rapid temperature increases such as observed in 1982/1983 and 1989/1990 (Fig. 3).

The inflow events and the associated variability in the ocean climate are major driving forces for the ecological conditions in the Barents Sea, affecting all trophic levels (Skjoldal and Rey, 1989; Blindheim and Skjoldal, 1992). The Barents Sea ecosystem is a zoogeographical transition zone where the inflowing Atlantic water is gradually transformed into Arctic water. This makes the Barents Sea strongly dependent on advective transport of boreal plankton with the inflowing water. The seasonal vertical migration of the dominant herbivore copepod *Calanus finmarchicus* interacts with the fluctuations in the ocean currents to determine the transport into the Barents Sea (Fig. 9). Atlantic water flowing into the Barents Sea during the late autumn and winter will contain few *C. finmarchicus*, which overwinters in deep water of the Norwegian Sea below the threshold depth at

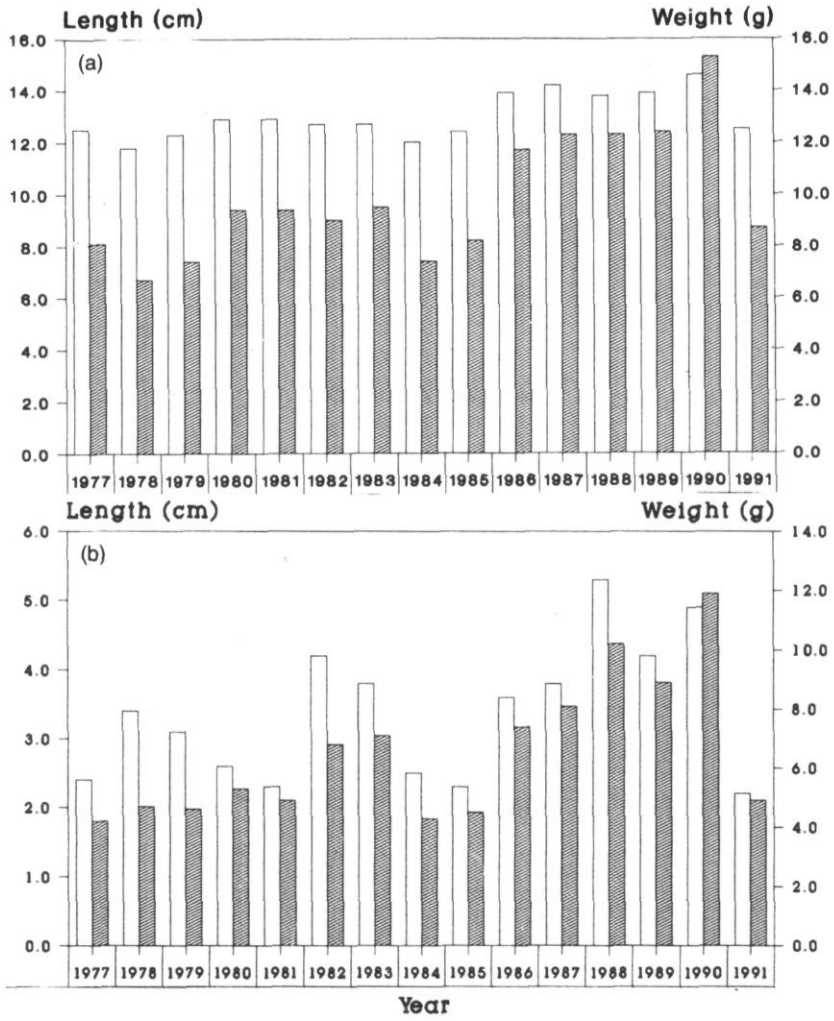

Figure 6. (a) Length and weight in September for two-year-old individuals, averaged for the total stock of Barents Sea capelin from 1977 to 1991. (b) Mean increase in length and weight in current year for the total stock of two-year-old capelin for the years 1977–1991.

the entrance to the Barents Sea. Water flowing into the Barents Sea during spring and summer will, on the other hand, contain the *Calanus* population, which then resides in the upper layer. This mechanism of low transport of *Calanus* during the overwintering period contributes to a lower and more variable zooplankton biomass in the southern and central Barents Sea than in the adjoining part of the Norwegian Sea (Degtereva, 1979; Skjoldal and Rey, 1989).

The major inflow event in 1982 apparently ended late in the winter 1982/1983. The southwestern Barents Sea was by then filled with Atlantic water that had entered during autumn and winter and that contained few *Calanus finmarchicus*. This is a likely explanation for the observed low biomass in the central Barents Sea in 1983 and 1984 (Skjoldal and Rey, 1989). We have previously shown that the zooplankton minimum was due primarily to a low abundance of *C. finmarchicus* (Skjoldal *et al.*, 1987).

Besides the influence of advective transport, there is also a large influence on zooplankton stocks by growth and predation within the Barents Sea. Most of the data in the time series of zooplankton biomass for subarea 7 (Fig. 4) are from Atlantic water, where *Calanus finmarchicus* is a dominant species. *C. finmarchicus* spawns in spring and the new cohort develops to reach the overwintering stage V by summer (Hassel, 1986; Skjoldal *et al.*, 1987). After the growth period, predation from capelin and other predators will reduce the abundance and biomass of *Calanus* during the autumn. Because of this cohort development and predation, a considerable seasonal variation in zooplankton biomass is expected. This is exemplified by results from June and August 1981 and from May and September 1987 and 1989 (Fig. 4).

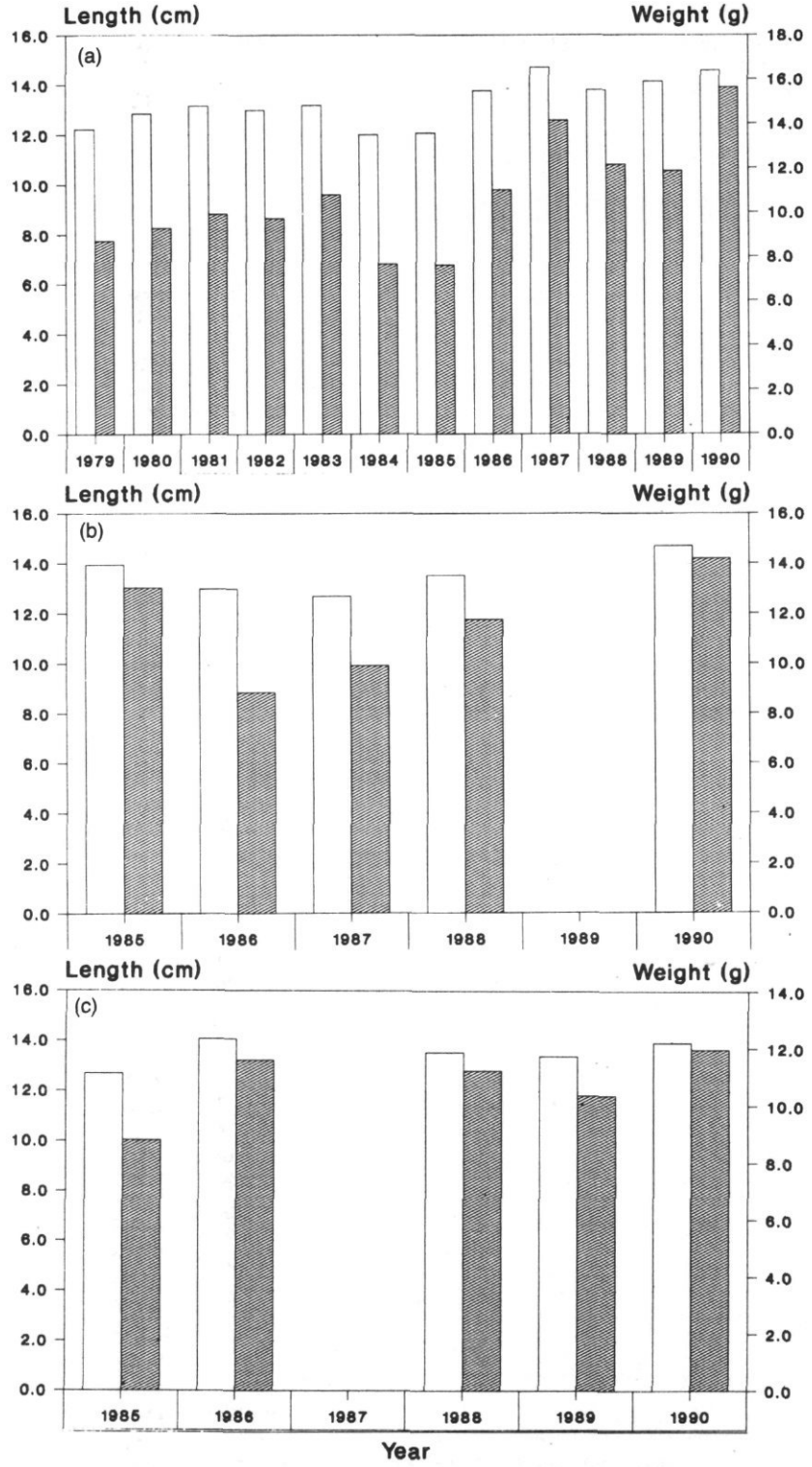

Figure 7. Mean length and weight in September of two-year-old capelin sampled in (a) subarea 7 from 1979 to 1990, (b) subarea 5, and (c) subarea 8 from 1985 to 1990.

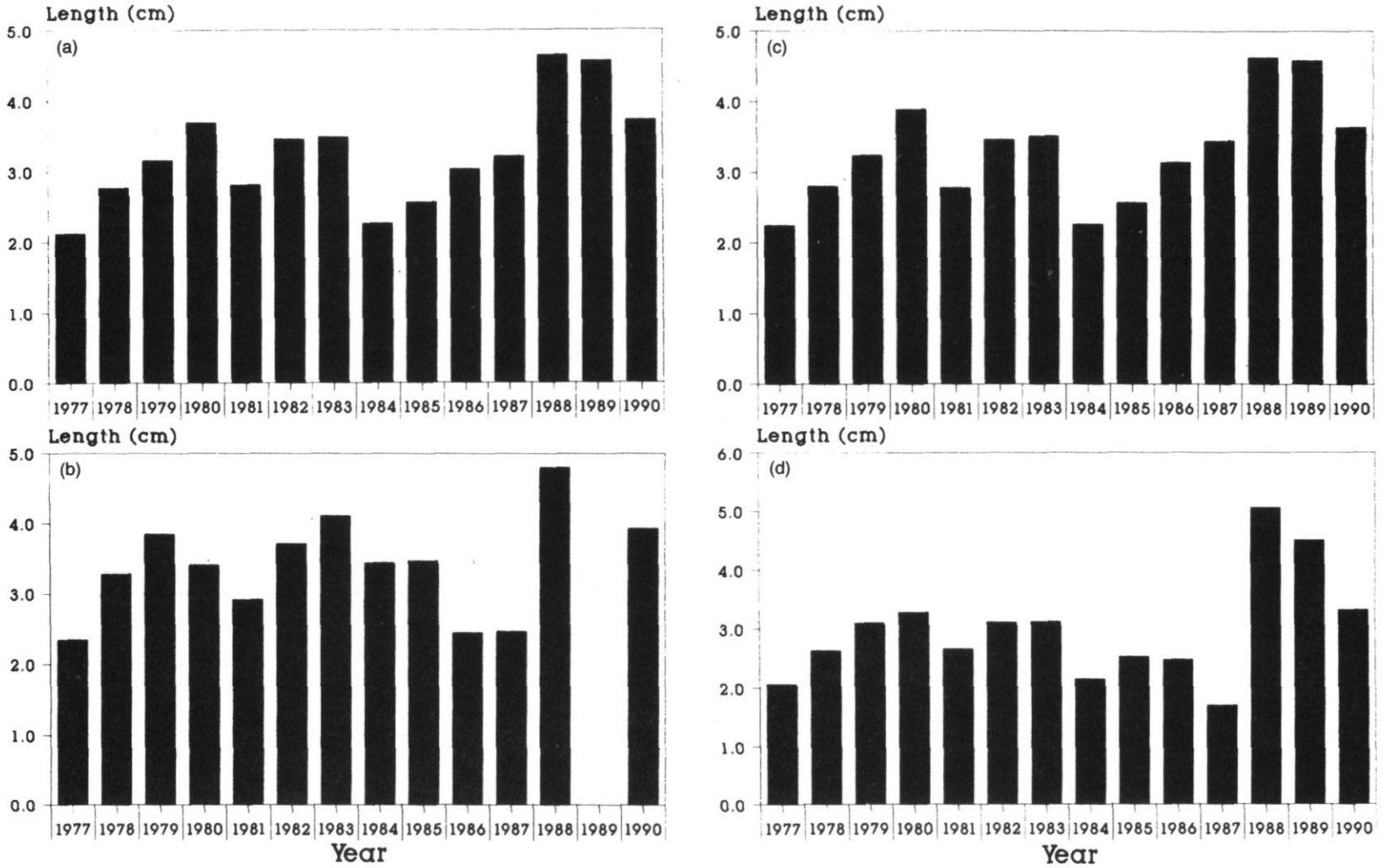

Figure 8. Estimated growth in current year as length increment calculated from otolith growth zones for capelin for the period 1977–1990. (a) Total area. (b) Subarea 5. (c) Subarea 7. (d) Subarea 8.

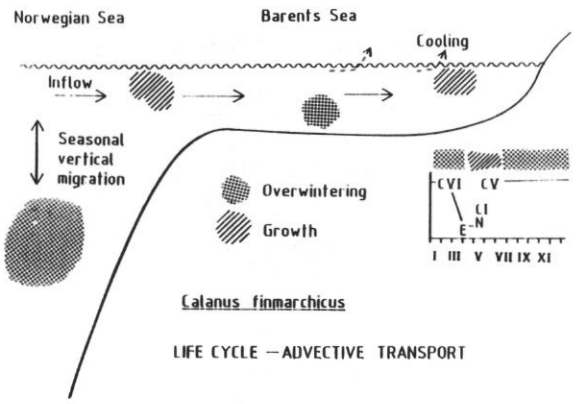

Figure 9. Life cycle and transport of *Calanus finmarchicus* into the Barents Sea. Animal behaviour, reflected in seasonal vertical migration to overwintering depth in the Norwegian Sea below the threshold depth at the entrance to the Barents Sea, interacts with Atlantic inflow to control the advective transport of *Calanus*.

The seasonal variation is one reason why a direct comparison between zooplankton biomass and capelin growth is difficult.

The growth of capelin has shown large variation during the late 1970s and the 1980s, with minima in 1977, 1984, and 1991 and maxima in 1980, 1982–1983 and 1988–1990 (Figs. 6 and 8). Comparing these results with the temperature and salinity anomalies in the Fugløya–Bjørnøya and Vardø-N sections (Fig. 3), it will be seen that the growth minimum in 1977 coincided with the onset of large negative anomalies. The low growth in 1984–1985 also appeared in a period of rapidly decreasing anomalies. The periods of high growth in 1980 and 1982–1983 occurred under rapidly increasing temperature anomalies. The period of high growth in 1988–1990 started under average temperature conditions followed by a rapid increase to high positive anomalies both in temperature and salinity. Consequently, there is a tendency for high growth to occur during periods of warming and low growth to occur during periods of cooling. This could reflect a general pattern of larger advective transport of plankton with the inflowing current during periods of warming and less transport during cooling periods following inflow events.

Temperature may also have a direct influence on growth and distribution of capelin. Gjøsæter and Loeng (1987) found that the mean temperature in the capelin feeding area varied from year to year during the period 1974 to 1985. In periods of increased inflow of warm Atlantic water, the capelin stock would partly compensate for the increased temperature by shifting the feeding area north- and eastwards. Correspondingly, there was a south- and westwards shift in distribution in periods of water cooling. Such shifts in distribution may act to reduce the effects of inflow events on growth rate.

Comparison of the temporal pattern of change in capelin growth rate (Figs. 6b and 8a) with the time series of zooplankton biomass (Fig. 4) reveals no significant correlation ($r = -0.24$, $n = 12$). A strict relationship between capelin growth rate and zooplankton biomass is not to be anticipated, however, for several reasons. One is the seasonal change in zooplankton biomass already mentioned. A second reason is the dynamical nature of the interaction between capelin and zooplankton, where capelin on its feeding migration can rapidly deplete the zooplankton stocks, particularly the preferred large food organisms distributed in the upper water layer (Hassel *et al.*, 1991). A third reason is the selection in sampling where large forms of zooplankton such as euphausids and amphipods are not representatively sampled by the vertical net hauls. Despite these limitations, there are some broad similarities between capelin growth rate and mean zooplankton biomass in the subareas. For subarea 7, both variables showed maximum in 1980, minimum in 1984, and maximum again in 1987 or 1988 (Figs. 4 and 8c). The patterns were different for subarea 5 in the southeastern Barents Sea, where capelin growth rate and zooplankton biomass were low in 1987 (Figs. 5a and 8b). In subarea 8 (the northeastern Barents Sea) there was no clear relationship (Figs. 5b and 8d).

The capelin stock size in numbers fluctuated by a factor of more than 200 during the period 1979–1991 (Fig. 10a). One would expect density-dependent influence on growth rate to be easily recognizable during such large stock size fluctuation. Such influence has for instance been shown for Atlantic herring where an inverse relationship between growth rate and year-class strength has been found (Toresen, 1986). However, the density dependency is complex in the present case of the Barents Sea capelin. Figure 10b depicts the growth history of three selected year classes. The 1978 year class was chosen as representative for the growth during a period of intermediate to high stock size from 1979 to 1982. By comparison, the 1982 year class showed considerably lower growth despite the rapid decrease in stock size from 1983 to 1986. We interpret this as an overriding influence by the low availability of zooplankton food on the capelin growth rate. The low growth rate was one factor which contributed to the rapid decline of the capelin stock. The event of inflow of Atlantic water and associated warming of the Barents Sea had thus a marked influence on capelin growth, but in an intricate and non-linear manner. Only through knowledge of the mechanisms involved (Fig. 9) can such relationships be properly explained in hindcast and used for predictive purposes in future models (Giske *et al.*, 1992).

The recovery of the capelin stock was surprisingly rapid from the low stock size in 1986 to a high level in 1990 (Hamre, 1991, 1992; Fig. 10a). The high individual growth rate (Figs. 6b and 8) and correspondingly early

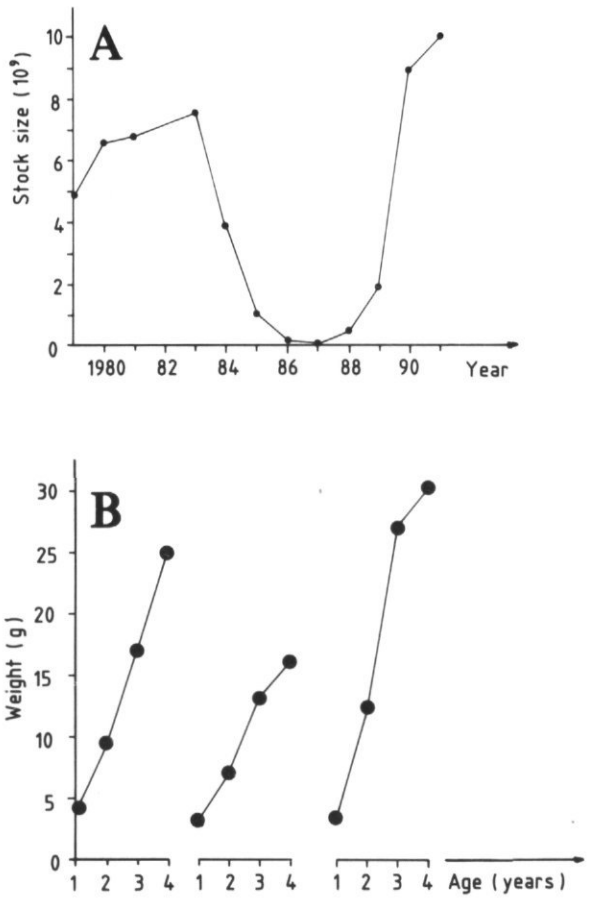

Figure 10. (A) Size of the Barents Sea capelin stock of two years of age or older for the period 1979–1991. (B) Growth curves as mean weights for 1 to 4 years old individuals for three selected year classes: 1978, 1982, and 1987.

maturation played an important role in this rapid recovery. The rapid growth of the 1987 year class is apparent from Figure 10B, which shows that the mean weight of three-year-old individuals of this year class was almost three times that of the 1982 year class. The highest capelin growth rates were seen in the period from 1988 to 1990 (Figs. 6b and 8). This corresponded to the second inflow event and warm ocean climate in 1989–1990. While the capelin stock was at its low level in the late 1980s, the Barents Sea ecosystem was considered to be in a disturbed state with disrupted energy flow to higher trophic levels (Borisov, 1991). As we see now, this situation was transient, and natural regulatory processes caused a rapid recovery of the capelin stock that rectified the "imbalance" of the ecosystem.

Reduced predation by capelin is likely to have contributed to a general increase in the amount of zooplankton after the minimum in 1984 (Fig. 4). This increase in zooplankton contributed in turn to high capelin growth rate. Feeding capelin selects larger zooplankters, and euphausids, amphipods, and large copepods such as *Calanus hyperboreus* and *C. glacialis* are important food items for the older year classes of capelin (Hassel, 1984; Panasenko, 1984). These large zooplankters have a life cycle of more than one year. Dominant euphausids in the Barents Sea, such as *Thysanoessa inermis* and *T. raschii,* reach maturity after two years (Dalpadado and Skjoldal, 1991). There will therefore be a delay in their population increases following a decrease in predation mortality. It is likely that the rapid decrease in capelin stock between 1984 and 1986 was associated with a marked reduction in predation and that the populations of euphausids and amphipods increased in response to this. There is indirect evidence of an increase in abundance of *Parathemisto* spp. from stomach content data for cod (Mehl, 1989, 1991). High abundance of large food items was probably a decisive factor for the rapid growth of capelin when the stock recovered from low levels.

The growth rate of capelin was again low in 1991 and comparable to growth rates in 1977 and 1981 (Fig. 6b). The large capelin stock size (Fig. 10A) is the probable cause for the reduced growth rate, through its effect on the feeding environment (Hassel *et al.*, 1991; Anon., 1992). The relationship between stock size and feeding conditions is complex, however, as both the total areal distribution and smaller scale distributional patterns will determine the available food levels and feeding conditions of capelin. The dynamic nature of the capelin–zooplankton interactions makes it necessary to address the issue of density-dependent growth more closely through the use of analytical and numerical models (Giske *et al.*, 1992).

In conclusion, we can state that variation in growth rate has played a decisive role in the large fluctuation of the capelin stock size in the recent decade. Large-scale changes in the physical conditions caused by the major inflow events and associated variability in ocean climate have been a primary driving force for the ecosystem dynamics (Sherman, 1991a, b). The physical forcing has, however, to a considerable degree acted through various biological interactions in the ecosystem. Thus we can link the major changes in capelin growth rate and stock size fluctuations with changes in zooplankton. Variation in fish recruitment has also had important effects on the ecosystem dynamics. Strong year classes of cod and herring in 1983 caused important interactions and effects in the system (Skjoldal and Rey, 1989; Hamre, 1991, 1992; Mehl, 1989, 1991). One probable effect was to cause recruitment failure of capelin in 1984 and 1985 due to predation on capelin larvae by juvenile herring (Fossum, 1992). The zooplankton minimum in 1984 may have contributed to low growth rate and thereby high predation mortality of capelin larvae in this period (Skjoldal and Rey, 1989).

Acknowledgments

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Recent variability in the fisheries of the North Atlantic

Jakob Jakobsson

Jakobsson, J. 1992. Recent variability in the fisheries of the North Atlantic. – ICES mar. Sci. Symp., 195: 291–315.

The trends in fisheries in six key areas of the North Atlantic during the period 1980–1989 are described and discussed. These are the Iceland–Greenland area, the fishing areas off Canada and the United States, the North Sea, the Baltic, and the Barents Sea. The Exclusive Economic Zone (EEZ) was extended to 200 miles just before the start of this period, resulting in a shift in the early 1980s from management by international organizations to management by coastal states. The different harvesting strategies which developed are discussed.

Exploitation rates continued at a high level in most fishing areas, especially in the two southernmost, the North Sea and off the US coast. Landings of demersal species in the North Sea continued at a high level during most of the period. By the late 1980s some important demersal stocks had been reduced to very low levels, which, with lower levels of recruitment, resulted in greatly reduced landings. The trend in the demersal fisheries in the US EEZ resembled that in the North Sea. Very high rates of exploitation and variations in recruitment have been associated with reduced abundance of the principal groundfish species, while other commercially less valuable species have flourished.

In the Barents Sea, Northeast Arctic cod suffered from a high rate of exploitation and a very low level of recruitment between 1976 and 1982, but estimates of the 1983–1985 year classes were high and rapid recovery was anticipated. However, forecasts of greatly increased yields did not materialize owing to the dramatic changes in the ecosystem during the latter half of the 1980s.

The variability in the demersal fisheries in the Iceland–Greenland area is reflected through fluctuations in the cod stock. During the 1980s recruitment to the Icelandic cod stock was very low except for the 1983 and 1984 year classes. The drift of a large proportion of the 1984 year class to Greenland waters stimulated the development of this fishery in the late 1980s.

Unlike other coastal states or management agencies, Canada declared a clear and definite management strategy at the end of the 1970s, i.e. to exploit demersal species, especially cod, at the low level of $F_{0.1}$. Stocks were expected to recover rapidly and yields to increase gradually towards the end of the 1980s. However, although the rate of exploitation of demersal species, especially the cod, was lower in Canadian waters than in most other areas it was nevertheless twice as high as the targeted fishing mortality. This discrepancy was largely attributable to a much lower recruitment to the major cod stock (northern cod) in the 1980s than anticipated.

One of the most striking common features in the North Atlantic during the 1980s is the recovery of many of the herring and mackerel stocks. During the late 1960s and early 1970s most pelagic stocks in the North Atlantic collapsed only to recover during the 1980s. The greatly increased level of recruitment to the North Sea herring during the 1980s compared with the 1970s is an outstanding example of this kind of extreme variability. Great fluctuations in abundance of the capelin stocks in the Northeast Atlantic took place during the 1980s. Thus, the Icelandic capelin stock collapsed and subsequently recovered during the first half of the decade, whereas the Barents Sea capelin stock went through the same cycle, collapsed in the mid-1980s, and showed clear signs of recovery at the end of the decade.

A common feature in the Northeast Atlantic is that the 1983 year class appears to have been especially strong in several fish species and fish stocks, both demersal and pelagic.

J. Jakobsson: Marine Research Institute, PO Box 1390, Skúlagata 4, Reykjavík, Iceland.

Introduction

In recent years the most dramatic variability in the total catch of fish has been in the collapse of many of the pelagic stocks, especially the herring stocks in the North Atlantic. The collapse of the Atlanto-Scandian herring has had the most drastic and long-lasting effect; practically all the other herring stocks have recovered. It is well known that a very important winter fishery has been based on the spawning migrations of the Atlanto-Scandian herring along the western coast of Norway for at least ten centuries. The periodicity of this fishery has been taken as an example of variations in stock abundance in earlier centuries, too. A discussion of these historical fluctuations is beyond the scope of this paper, but until this century the winter fishery along the west coast of Norway was conducted by beach seines only and was therefore entirely dependent on the herring spawning right on the beach of the Norwegian west coast. If in earlier centuries the herring had periodically spawned in deeper waters (as is the case in recent decades), the fishery would have failed. Therefore wrong conclusions on stock abundance, based on fluctuations in fisheries alone, can easily be reached, especially if the limitations of the fishing gear used in earlier centuries are not taken into consideration.

Although the winter fishery along the west coast of Norway continued to be the most important one on the Atlanto-Scandian herring, another important fishery based on this herring stock developed at the beginning of the century as a consequence of the introduction of the purse-seine in the North Atlantic. This was the North Icelandic herring fishery, which was based on feeding migrations of the Atlanto-Scandian herring. However, during the first half of the 20th century these two fisheries developed independently and it was not until the theories of the Icelandic biologist Árni Friðriksson on the transatlantic migrations of the Atlanto-Scandian herring had been proven, through tagging operations (the first trans-oceanic tags were recovered in 1949), that the herring industry realized these two fisheries were based on the same herring stock. In response to a proposal by the Nordic countries, ICES initiated preseasonal herring/environmental surveys which started in 1952 and continued until the collapse of the Atlanto-Scandian herring in the late 1960s. The purpose of these surveys was to chart the northwest feeding migration of the Atlanto-Scandian herring towards the Iceland–Jan Mayen area. By the beginning of the 1960s these preseasonal surveys had gained much popularity because they had led to the extension of the summer fishing season (Fig. 1). The herring fleet therefore waited with great expectations for any news from the survey vessels during each spring. This was also the case in 1965. The survey started on the 5th of May. On approaching north Iceland it had become clear that survey conditions were completely different from previous years (Fig. 2). The extension of the sea ice was much greater than had been observed since the cold year of 1918, the sea temperature was very low, and there was hardly any sign of the important food species *Calanus finmarchicus* in the area west and north of Iceland. When coming to the area of the East Icelandic Current, temperatures below 0°C were common and on completion of the standard hydrographic sections northeast and east of Iceland there was no sign of the Polar Front. The hydrographic sections had therefore to be extended much farther east than in any previous year of these surveys in order to find the Polar Front on the east side of the East Icelandic Current.

This was in fact the beginning of the "Great Salinity Anomaly of the 1970s", which culminated in the Iceland Sea in 1968–1969 (Fig. 3). The immediate effect on the food chain and the herring migration pattern was obvious. Because of stratification of the surface layers the phytoplankton bloom lasted for only a very short period each spring. The populations of *C. finmarchicus*, the most important food species for the herring, disappeared almost completely. The highly productive area north of Iceland, which had sustained enough food for the largest herring stock in the world, was like a desert. These environmental changes had a dramatic effect on the migration pattern of the herring, which now undertook a very large and extensive migration all the way to Jan Mayen and Spitzbergen in search of food. These dramatic environmental changes and their effect on the herring stock are often forgotten when the reasons for the collapse of the Atlanto-Scandian herring are under consideration.

Although the ice-years north of Iceland from 1965 to 1970 had an immediate and drastic effect on the herring migrations (Jakobsson, 1980b) it is becoming more and more obvious that these events also had long-lasting effects on the distribution of the watermasses in the North Atlantic (Malmberg, 1969). Recently, Dickson, *et al.* (1988) traced the movement of this "slug" of fresh water which has freshened much of the upper northern North Atlantic during the past 25 years. This phenomenon is further analysed by Aagaard and Carmack (1989). The "Great Salinity Anomaly" has probably generated more variability in fisheries during the last quarter of a century than any other hydrographic event in recent years.

Following the work of Johan Hjort (1914) on the fluctuations in fish populations, marine scientists have tried to explain all fluctuations in terms of variability in recruitment. When marine scientists acquired the tools with which to study the effect of fisheries on the fish populations, the tendency was to forget about the natural effect of changing environmental conditions and to explain all fish stock fluctuation by fishing activities. To gain a better understanding of the variability in the oceans it is important to analyse and understand the effect of both natural causes and human activities.

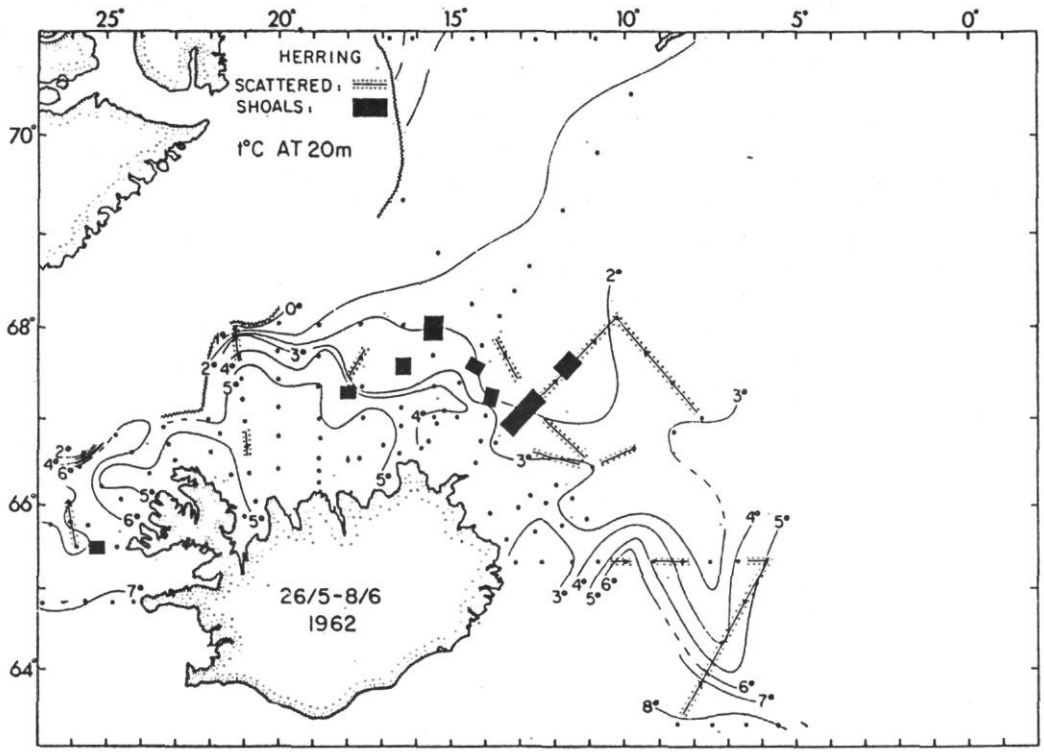

Figure 1. Distribution of herring schools and temperature at 20 m depth off Iceland in May–June 1962. From Jakobsson (1980b).

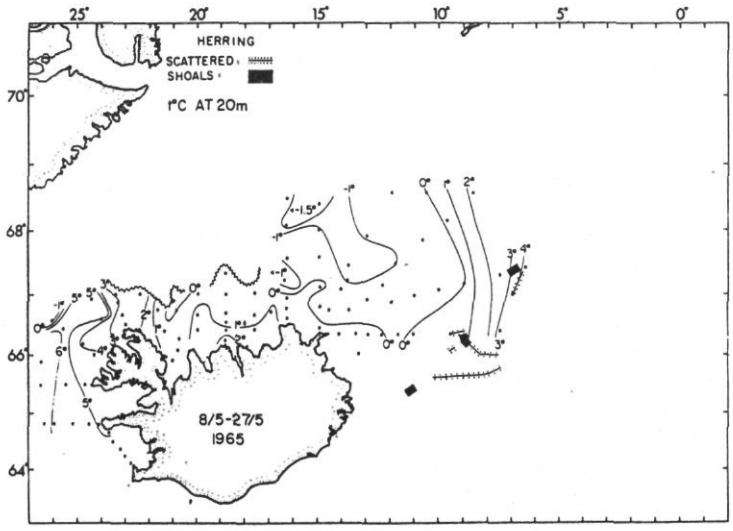

Figure 2. Distribution of herring schools and temperature at 20 m depth off Iceland, May 1965. From Jakobsson (1980b).

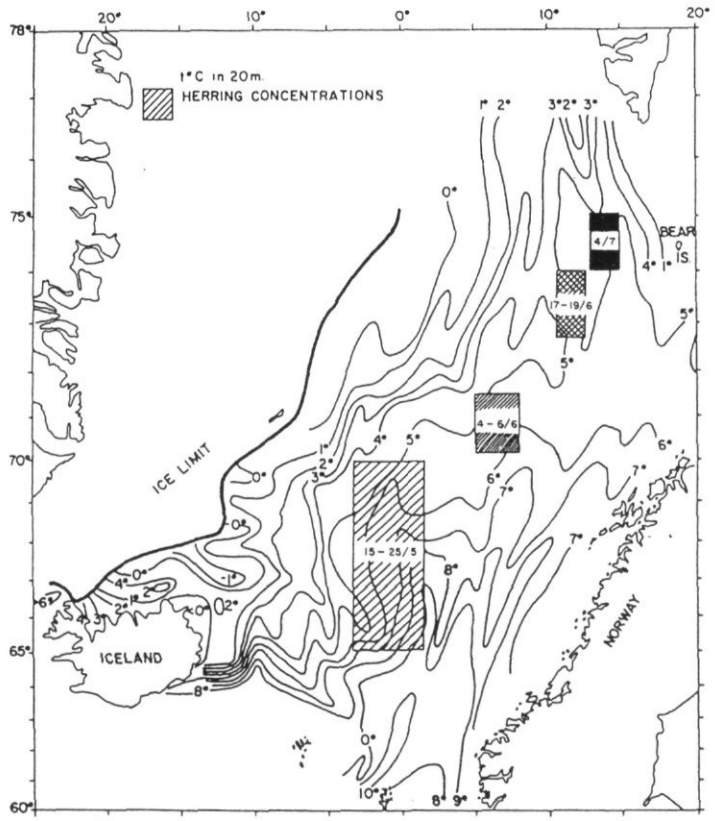

Figure 3. Temperature at 20 m depth (May–June) and herring migrations (northwards), May–July 1968. From Jakobsson (1980b).

In the following sections recent variability in fisheries will be examined in a few important fisheries distributed across the North Atlantic. In doing so the path of the "Great Salinity Anomaly" will be followed as far as possible.

The cod fishery at West Greenland

The rise and collapse of the cod fishery at West Greenland 1930–1970 (Fig. 4) has been associated with climatic changes (Hovgård and Buch, 1990). Although the cod stock at West Greenland is probably more dependent on climatic variations than most other fish stocks it is worth noting that recruitment was drastically reduced as early as 1964, when the stock was under very great fishing pressure, i.e. just before the change in marine climate took place. The peak effect of the "Great Salinity Anomaly" on Iceland to West Greenland waters was during 1969–1972, although it probably had an influence on these waters as early as the mid-1960s (Blindheim, 1967; Malmberg and Svansson, 1982). The presence of the "Great Salinity Anomaly" during the late 1960s and early 1970s coincided with an almost complete recruitment failure in the West Greenland cod stock (Fig. 5). Temporary recoveries have been due to drifting of 0-group fish from Iceland, especially the 1973 and 1984 year classes (Vilhjálmsson and Magnússon, 1984), but a strong year class, as seen before 1964, has not been produced at West Greenland in the past 36 years. The collapse of this cod stock is an example of the combined effect of overfishing and unfavourable environmental conditions.

The most dramatic event in this fishery during the 1980s was the large-scale drift of 0-group cod from Iceland in the autumn of 1984 (Fig. 6). During the spring

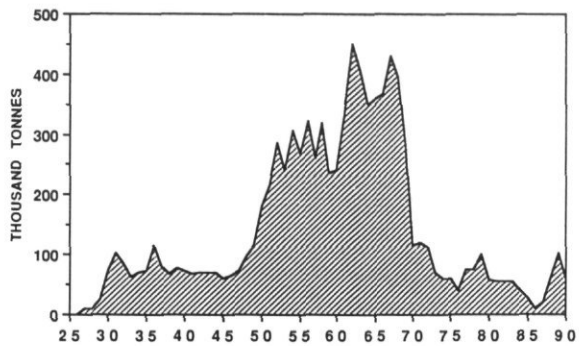

Figure 4. Annual cod catches off West Greenland, 1925–1990. Redrawn from Hovgård and Buch (1990).

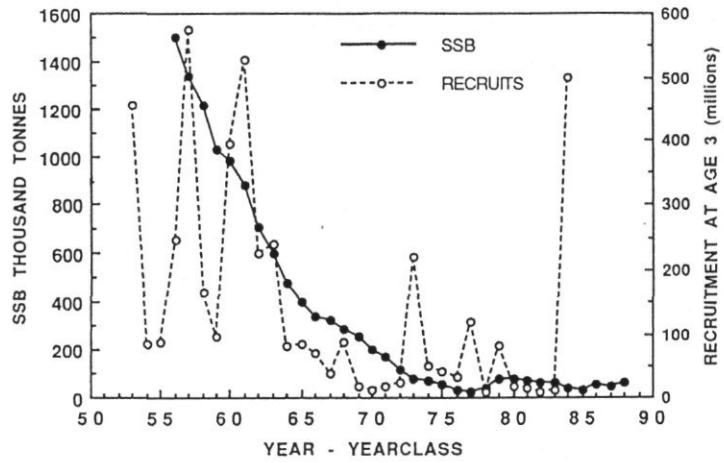

Figure 5. Spawning-stock biomass and recruitment in West Greenland cod, 1953–1988. The estimate of the 1984 year class is based on results from groundfish surveys, whereas estimates of other year classes are derived from VPA. Redrawn from Hovgård and Buch (1990).

of 1984 the sea temperature west and north of Iceland was above normal (Fig. 7). The Atlantic water had brought the cod larvae north along the west coast of Iceland and their arrival on the nursery grounds north of Iceland was earlier than usual. However, during the August 0-group survey it became evident that the Denmark Strait had been blocked by a tongue of cold Arctic water blocking the Irminger water (Fig. 8), and thereby also 0-group cod, from following its normal path into the Iceland Sea. Instead it was forced towards Greenland. Repeated surveys later in the same year showed that this tongue of Arctic water had disappeared from the strait and the flow of Irminger water to the Iceland Sea was normal again. The development of such cold Arctic tongues across the Denmark Strait is not uncommon. One in September 1964 was probably the first indication of the "Great Salinity Anomaly" which developed in the Iceland Sea during the late 1960s. The drift of 0-group cod to Greenland waters in 1984, and also to some extent in 1985, resulted in a temporary increase in the cod

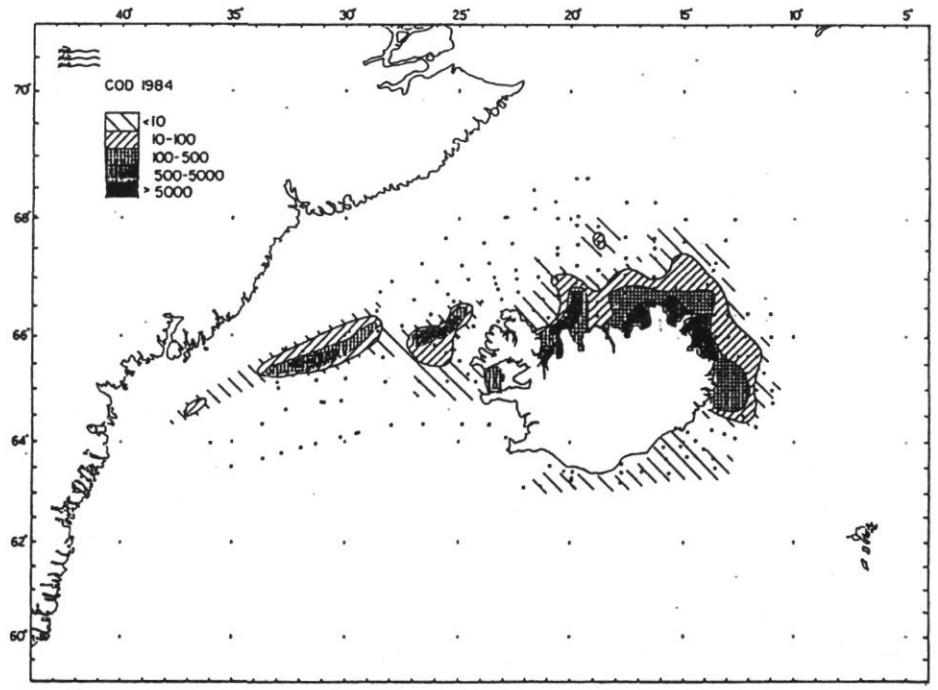

Figure 6. Distribution pattern of 0-group cod, August 1984. From Vilhjálmsson and Magnússon (1984).

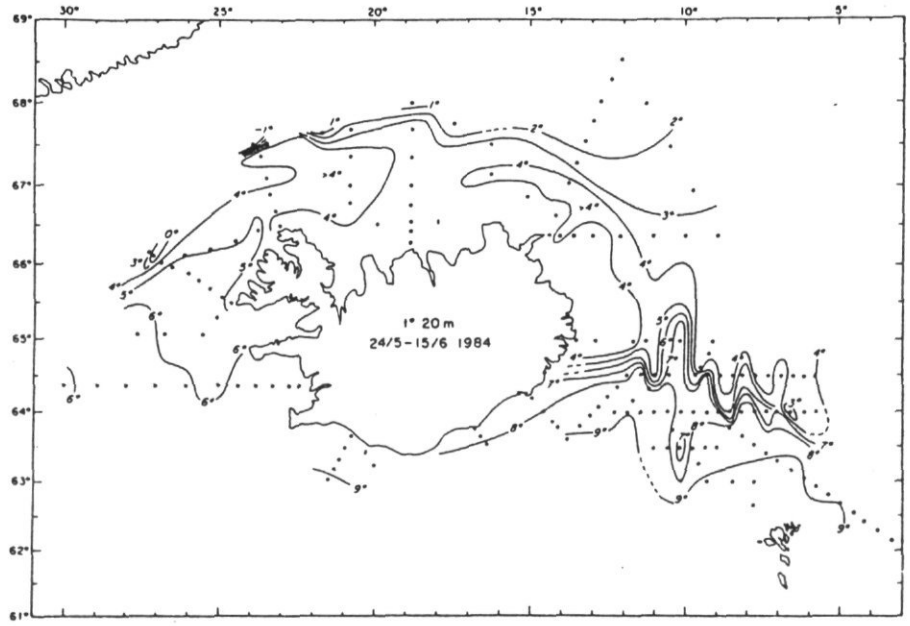

Figure 7. Temperature at 20 m depth, May/June 1984. Data provided by the Mar. Res. Inst. Reykjavík.

fishery at West Greenland. However, because of the very low growth rate of these year classes in Greenland waters the increase in the fishery was more in number than in weight. This was only a temporary increase because by 1990 the cod had almost disappeared from Greenland waters.

The fisheries in Canadian waters

Total landings of all species from Canadian waters during the 1980s amounted to about one-and-a-half million tonnes. Most of the commercially important marine species in Canadian waters are divided into

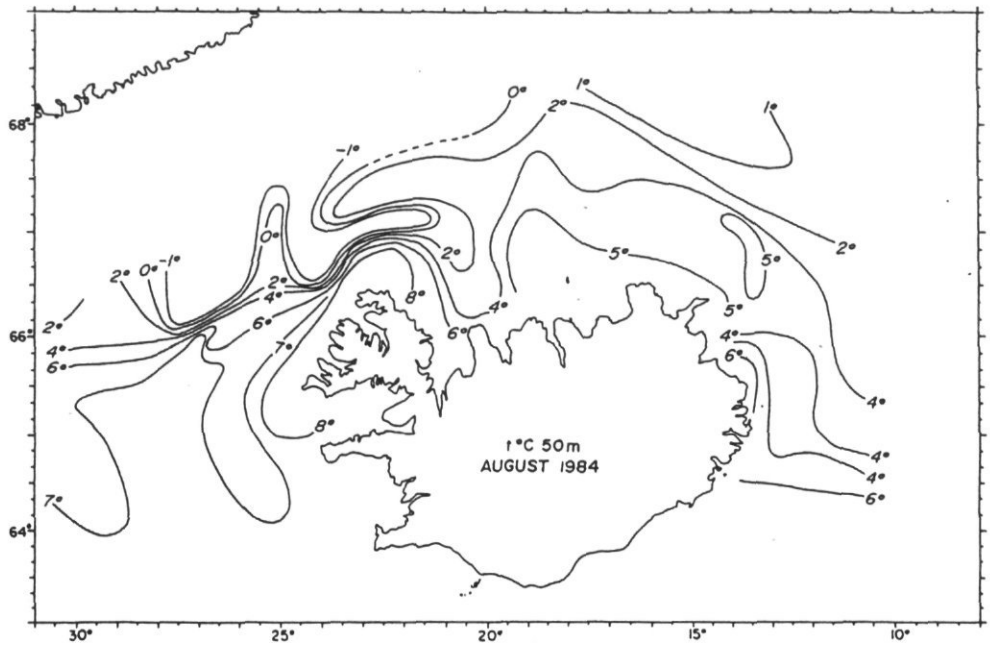

Figure 8. Temperature at 50 m depth, August 1984. Data provided by the Mar. Res. Inst. Reykjavík.

several stocks. Thus there are no fewer than 10 cod stocks and about the same number of herring stocks.

Total landings of cod from Canadian waters during the 1980s were remarkably steady at between 500 000 and 600 000 tonnes (Fig. 9). During the 1960s cod landings ranged from 900 000 to 1 400 000 tonnes. Subsequently, in the 1970s, there was a rapid decline.

The main source of variability in the total landings has come from the northern cod off Labrador and eastern Newfoundland as well as the cod stock on the Grand Banks. During the 1950s the catches of the northern cod fluctuated between 200 000 and 300 000 tonnes. In the 1960s there was a gradual increase in the catches, with a peak in 1969 of 850 000 tonnes. This was followed by a rapid decline to a minimum of 171 000 tonnes in 1979. Thereafter, catches increased to about 230 000 tonnes in 1983 and remained at about that level during the rest of the decade. Similar variability is observed on a smaller scale for the cod stock on the Grand Banks; the highest landing occurred in 1967, followed by a gradual decrease until 1978 when only 15 000 tonnes were landed. During the 1980s landings have varied between 30 000 and 50 000 tonnes.

It is generally accepted that the increase in yield from the northern cod stock during the 1960s was due to greatly increased fishing effort and resulting increased fishing mortality (Fig. 10). In addition, recruitment was at a high level during the early 1960s (Fig. 11). Some doubts have been expressed concerning the validity of the recruitment figures from this period (Pinhorn and Halliday, 1990), which may be biased owing to misreporting and insufficient sampling of the catches. The fact remains, however, that it would have been impossible to keep up the high catches for 15 years without a high level of recruitment during at least part of this period. It is worth noting that there was a sudden decline in the level of year-class strength from about 600 million age 3 cod in 1968 to 253 million in 1969, and from then until 1972 recruitment was at an exceedingly low level (Fig. 11).

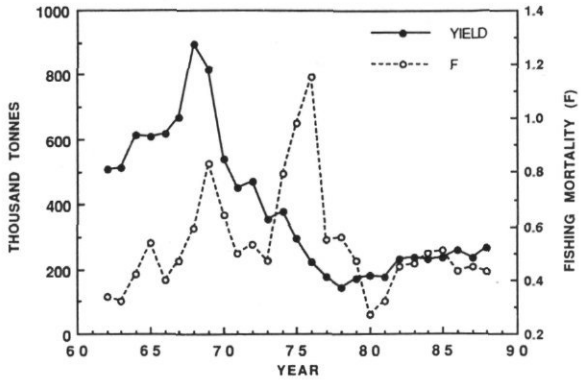

Figure 10. Yield and fishing mortality in the northern cod stock, 1962–1988. Data from the CAFSAC 1989 report.

The combined effect of this change in recruitment level, as well as the very high exploitation rate until the extension of the fishing jurisdiction in 1978, brought about the rapid decline in stock abundance during the 1970s.

When the Canadian government extended the fishing jurisdiction to 200 nautical miles the northern cod stock was at a very low level. At the time a programme was initiated to rebuild the fish stocks off the east coast of Canada and subsequently greatly increased yields were expected. The focal point in this programme was to reduce the fishing mortality to what is called $F_{0.1}$. For the northern cod stock this corresponds to a fishing mortality of F = 0.2, which is about equal to the assumed natural mortality. This is only about one-quarter of the fishing mortality on the stock during the 1970s. To realize this target a total allowable catch was set. If achieved, forecasts showed that stocks would recover rapidly and that yields would increase as a result of recovering stocks. Figure 12 shows that a large part of the expected increase in yields was to come from the northern cod stock (Anon., 1982). The expected catches

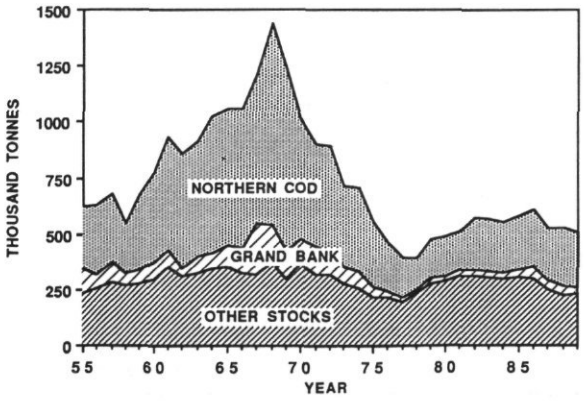

Figure 9. Annual landings from Canadian cod stocks, 1955–1989. Data from the CAFSAC 1989 report.

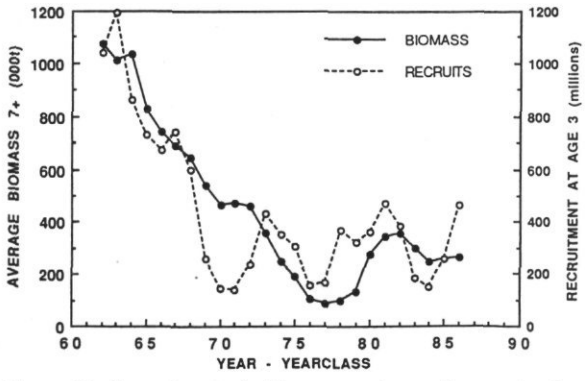

Figure 11. Spawning-stock biomass and recruitment in the northern cod stock, 1962–1986. Data from the CAFSAC 1989 report.

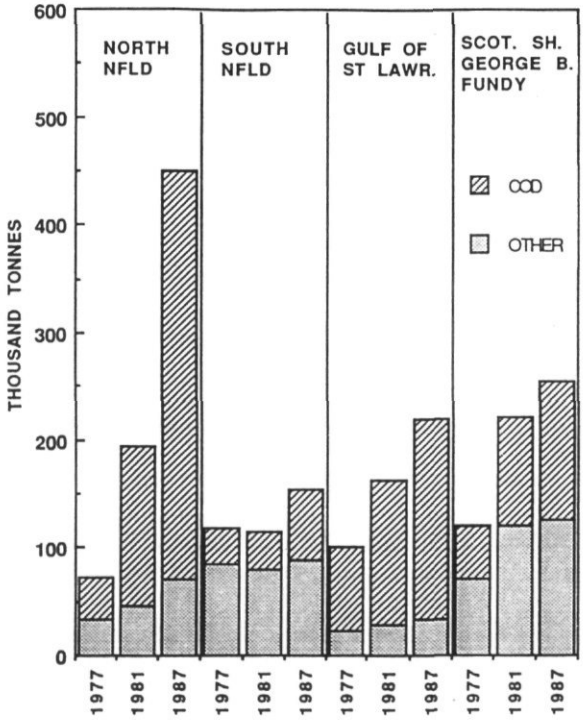

Figure 12. Regional breakdown of Canadian catches of groundfish for the years 1977 and 1981. Prognosis for 1987. Redrawn from Anon. (1982).

in 1987 were about 450 000 tonnes instead of the 256 000 tonnes which were in fact caught. In the projections, which showed the rapid recovery of the stock, the scientists used the historical average annual level of recruitment, i.e. corresponding to about 600 million three-year-olds. In retrospect, it can be seen that the average annual recruitment during the period from 1969 to 1985 was 276 million three-year-old cod, which is less than half the recruitment used in the projections. Since the sustainable yield of any fish stock is directly proportional to the recruitment, it is understandable that the projections did not materialize. It was not until the very end of the 1980s, however, that it was realized something was going wrong in the rebuilding of the northern cod. It now appeared that the total stock size was much smaller than had been predicted from all previous estimates. The main finding of the 1989 assessment (Anon., 1990a) was that while the stock had not declined during the 1980s it had not grown at a sufficiently rapid rate to justify the expectations. It was further concluded that the fishery during the 1980s had resulted in a fishing mortality which was more than twice the expected fishing mortality of $F = 0.2$. Thus, the fishing mortality had been underestimated throughout the 1980s by more than 50%. The underlying reason for the wrong assessment for such a long time was the assumption of a recruitment level more than twice that actually experienced. It has been concluded (Harris *et al.*, 1990) that the main reason for the low recruitment was the low level of the spawning stock. If this is true, then it is clear that the northern cod is the first stock of cod in the world where recruit-overfishing has been demonstrated, i.e. the northern cod has been so depleted that the spawning stock is unable to produce average or normal numbers of progeny. Figures 11 and 13 show the stock–recruitment relationship. The weakness in the above theory of recruit-overfishing is that the stock has not started to produce strong year classes again. Therefore, the strong stock–recruitment relationship could possibly be a temporal trend like the one observed in the North Sea herring in the 1970s. The fact remains, however, that the northern cod has not produced a large year class comparable with those of the 1960s during a period of 20 years. Koslow and Thompson (1987) concluded that there was some evidence that stock reproductive output was not significantly related to year-class success, arguing that it tended to be associated with large-scale meteorological patterns and offshore winds. In the middle of the 1980s, when Koslow and Thompson carried out their analyses, the assessment of the recovery of the northern cod was very different from that of today.

On the possibility of environmental factors affecting recruitment Harris *et al.* (1990) made the following remark: "In the course of its public hearings as well as in oral and written submissions, the panel was repeatedly alerted to the widely held belief that water temperature was an important controlling factor in some aspects of the northern cod life cycle and in respect of the availability or otherwise of fish to gear at specific times and locations. This view was shared by fishermen and scientists alike. Unfortunately, support for this contention rests for the most part on anecdotal or intuitive evidence. That in itself does not refute its potential validity but it does put the burden of proof on the scientist to define such a relationship if one in fact does exist." The

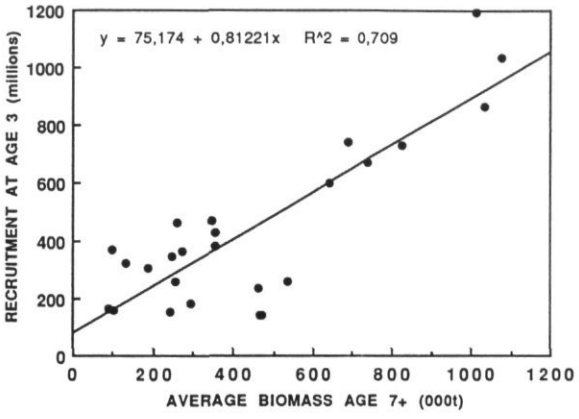

Figure 13. Stock and recruitment relationship in the northern cod stock, 1962–1986. Data from the CAFSAC 1989 report.

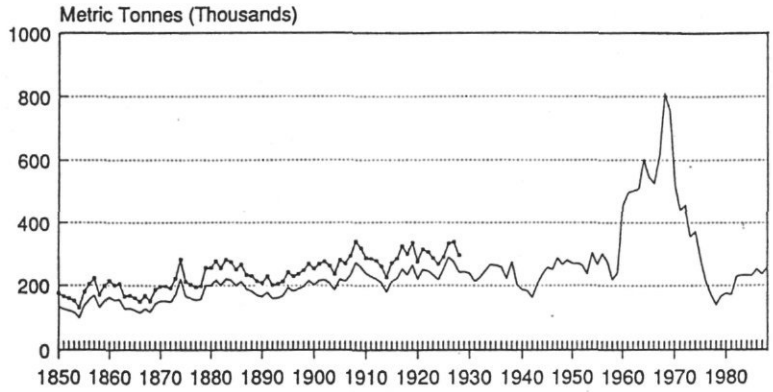

Figure 14. Annual landings from the northern cod stock, 1850–1989. From Harris *et al.* (1990).

report continues: "In conclusion, it is apparent that the distribution, survival and behaviour of the northern cod are influenced by a variety of oceanographic parameters and processes. These are, however, at present poorly understood". Thus Harris *et al.* (1990) concluded that the cause of the greatly reduced abundance of the northern cod and the lack of rapid recovery was fishery induced and if there were oceanographic factors which could possibly play some role in this, then they were not understood. In reaching this conclusion Harris *et al.* (1990) appear to have overlooked the possible effect of the "Great Salinity Anomaly".

The stock–recruitment relationship illustrated here is unusual because most cod stocks are very resilient to high rates of exploitation, i.e. high levels of fishing mortalities. This is primarily because cod stocks appear to be able to produce normal numbers of recruits despite low levels of spawning stock. If, however, during a period of high fishing mortality there is a sudden change in the recruitment level it is clear that these two factors will rapidly bring the stock down. This is exactly what happened in the case of the northern cod at the beginning of the 1970s. The "Great Salinity Anomaly" was passing from West Greenland waters to Labrador waters just at this time (Dickson *et al.*, 1988). And the spawning grounds of the northern cod are on the outer slopes of the Labrador and Newfoundland continental shelf where the signal from the "Great Salinity Anomaly" was strongest. On the other hand recruitment to other cod stocks which spawn in coastal waters was not affected to such a degree, their yield remaining more or less constant (Fig. 9). Figure 11 shows that there was a temporary recovery in the recruitment to the northern cod stock when the "Great Salinity Anomaly" had passed from Canadian waters in 1973.

In these discussions it has been taken for granted that the high yields and high levels of recruitment during the 1960s and early 1970s were normal for the northern cod stock. Figure 14, however, shows that the yield of the northern cod stock fluctuated without trend between 200 000 and 300 000 tonnes from 1950 to 1960. The yield during the 1980s was on the same level. Indeed, it is only from 1960 to 1972 that there was a much higher yield. The question therefore remains whether the 1960s were unusual in producing a short-term outburst of cod in this area due to favourable environmental conditions.

Fisheries off the Northeastern United States with special reference to Georges Bank

The total landings from this area, about half a million tonnes in recent years, are based on no fewer than 33 species, some of which are divided into several stocks. In the report "Status of the Fishery Resources off the Northeastern United States" (Anon., 1989b) the reader is aided by the grouping of the stocks and species into six categories, the first being the Principal Groundfish Species. This group has in recent years accounted for just under 30% of the total landings. The second group is the Principal Pelagic Species, which in the same period accounted for 24% of the total landings. The third group, the Invertebrates, accounts for about the same proportion. The fourth group is Other Pelagic Species and accounts for about 10% of the catch. Finally come Flounders and Other Groundfish with smaller proportions.

Figure 15 shows the total landings of the most important groundfish species (cod, haddock, yellowtail flounder, and redfish) from off the Northeast USA coast during the period 1925–1990. There are four peaks – about 1930, during the early 1940s, in the mid-1960s and, finally, a minor peak during the early 1980s. This was followed by a rapid decline during the late 1980s. Similarly, there has been a sharp decline in c.p.u.e. and to a lesser degree in the survey index for these species during the late 1980s (Fig. 16).

The landings from the Georges Bank cod stock (Fig. 17) during the period 1962–1988 fluctuated without a

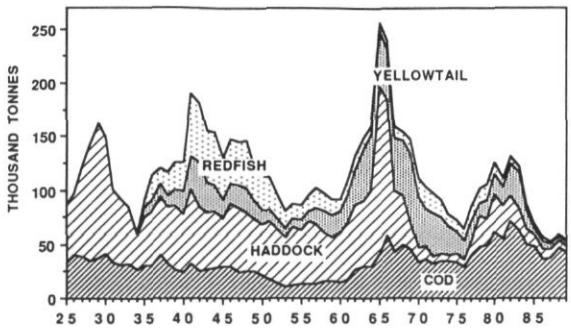

Figure 15. Annual total landings of cod, haddock, yellowtail flounder, and redfish from off the northeast coast of the USA, 1925–1989. Data provided by Northeast Fisheries Centre, Woods Hole.

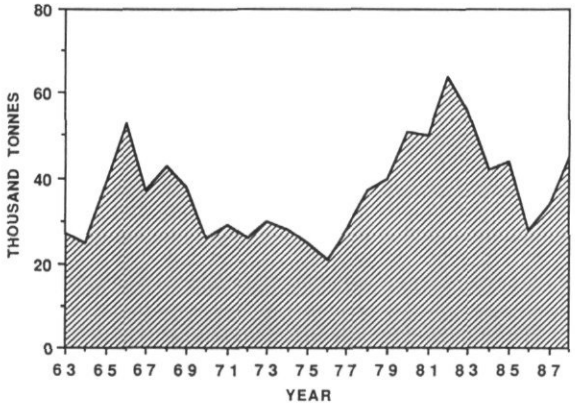

Figure 17. Annual landings of Georges Bank cod, 1963–1988. Data from same source as in Figure 16.

trend with two peaks, one in the mid-1960s and one in the early 1980s. The long-term potential catch from this stock has been estimated to be about 35 000 tonnes. This is similar to the level harvested in the last few years. However, concern has been expressed about the rapid decline of the spawning stock (Anon., 1989b). In recent years the fishing mortality has reached record heights ($F = 0.95$ for ages 3+), more than three times F_{max}. Towards the end of the 1980s there were some indications of increasing stock abundance and spawning stock biomass as a result of two strong year classes from 1985 and 1987 (Fig. 18). Despite the very high exploitation rate in the most recent years there is no sign of reduced recruitment level. Therefore the stock is probably growth overfished, but not yet recruitment overfished.

The landings of Georges Bank haddock during the period 1930–1989 are shown in Figure 19. In 1965 the catch of Georges Bank haddock jumped to 150 000 tonnes, three times that caught previously, before declining rapidly to a very low level. Catches increased temporarily from 1977 to 1980, reaching 28 000 tonnes, but declined thereafter to 6000 tonnes in 1987. The stock suffered from a succession of weak year classes during the period 1979–1987 with the exception of the 1983 and 1985 year classes (Fig. 20). However, fishing effort, mortality rate, and discards have been high in recent years and appear to have reduced the 1983 and 1985 year classes rapidly. As in other depleted stocks, catch levels are primarily dependent on the recruiting year class. The present catch level of less than 10 000 tonnes is only about 20% of the long-term potential yield. It is clear that following a very high level of recruitment at the beginning of the 1960s the stock has suffered from recruitment failure since 1964. Any recovery during the 1980s has been squashed by overfishing.

Not all the demersal species in the area are overfished. The proportion of skate and ray has increased sharply in the catches of the trawl surveys carried out on Georges Bank, where survey catches of skate and ray have changed from roughly 25% by weight in 1963 to nearly 75% of the groundfish catches in recent years. Such large increases in the relative biomass of very low-valued species has raised concerns about the possible negative

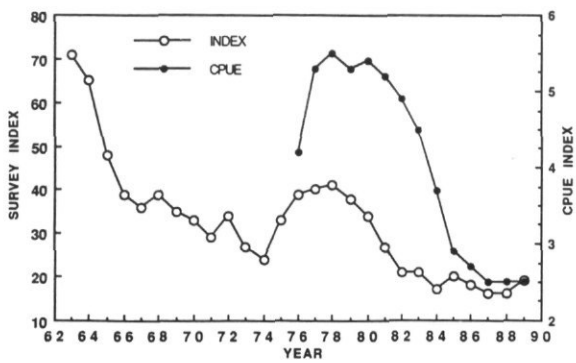

Figure 16. USA survey and c.p.u.e. abundance indices for principal groundfish and flounders. Data from Anon. (1989b).

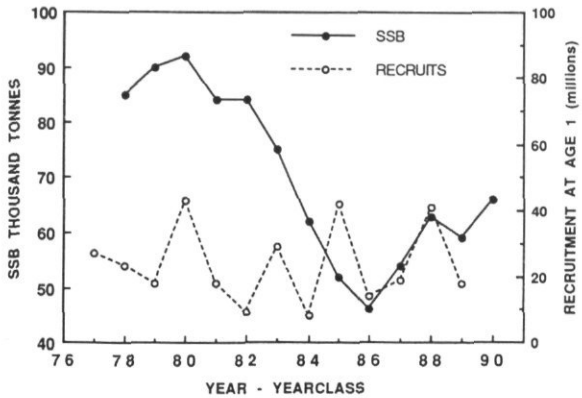

Figure 18. Spawning-stock biomass and recruitment in Georges Bank cod, 1977–1990. Data from same source as in Figure 16.

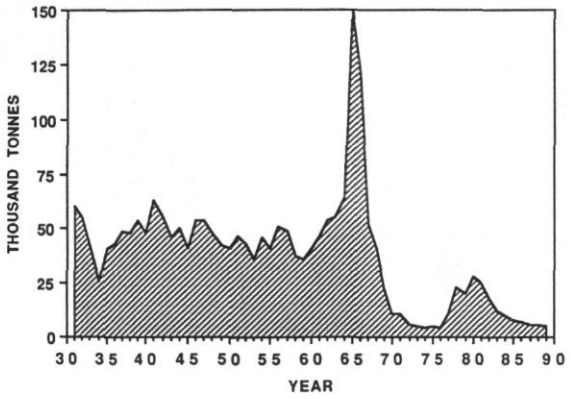

Figure 19. Annual landings of Georges Bank haddock, 1931–1989. Data provided by Northeast Fisheries Centre, Woods, Hole.

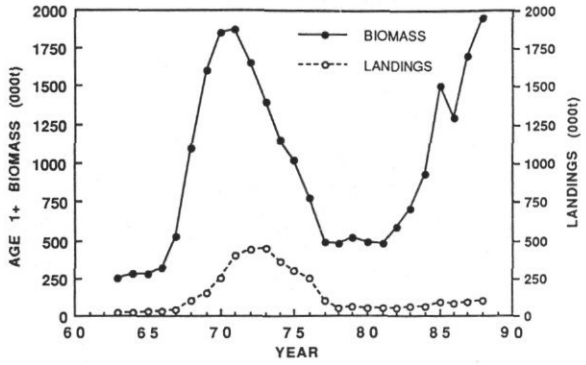

Figure 21. Stock abundance and landings of Atlantic mackerel in the area Labrador to North Carolina, 1963–1988. Data from Anon. (1989b).

effects of these predators on overall fishery yields (Anon., 1989b).

The landings and stock biomass of Atlantic mackerel during the period 1963–1988 are shown in Figure 21. The mackerel landings increased dramatically during the late 1960s, reaching a peak of roughly 400 000 tonnes in 1973. Landings subsequently declined to about 30 000 tonnes in the late 1970s but have increased somewhat during the 1980s. The 1975–1980 year classes were all relatively weak. During the 1980s the recruitment level appears to have increased and this has been the reason for the rapid recovery of the stock, as shown in Figure 10. Thus, the total stock biomass was almost two million tonnes at the end of the 1980s and has reached the same level as at the beginning of the 1970s. At present the stock is under-exploited.

Following the collapse of the Georges Bank herring there has been no directed fishery on this stock during the last 15 years. However, clear indications of recovery of the stock have been visible during the late 1980s and prospects for redevelopment of the fishery are being considered.

Georges Bank is one of the most productive marine ecosystems in the world, with a primary production of up to 450 g C m^{-2} $year^{-1}$ (O'Reilly *et al.*, 1987). With its proximity it is also one of the best-studied large marine ecosystems in the world (Backus, 1987). Of special interest has been study of the recruitment mechanisms on Georges Bank, in particular in relation to (a) the availability of prey during the larval period and (b) the physical transport of eggs and larvae to unfavourable locations (Cohen *et al.*, 1988). Although starvation mortality does not appear to be population limiting over the range of prey densities observed above the thermocline, storm events have sometimes disrupted stratification and may possibly have caused starvation. Eggs and larvae may also be advected off the bank during storms or as a result of so-called "warm core rings" which spin

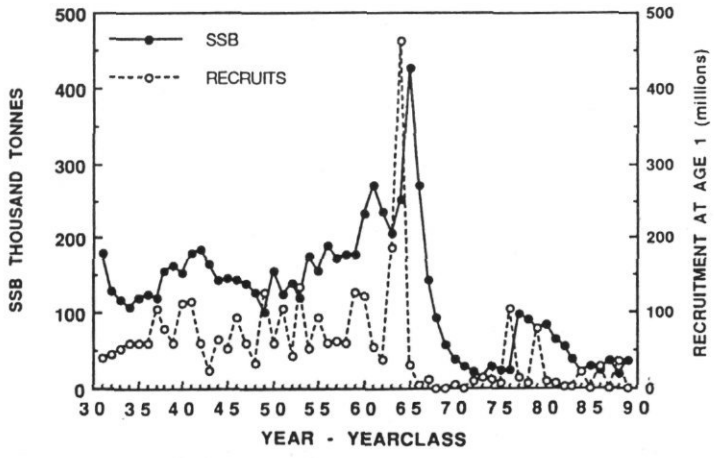

Figure 20. Spawning-stock biomass and recruitment in Georges Bank haddock, 1931–1989. Data provided by Northeast Fisheries Centre, Woods Hole.

off from the Gulf Stream. Several investigations have reported that mackerel and herring predation may have a major influence on the survival of 0-group cod, haddock, and sand lance (e.g. Michaels, 1991; Fogarty *et al.*, 1991; Sissenwine, 1986). Furthermore, abnormally low temperatures may threaten the survival of haddock eggs and larvae in some years.

Despite extensive research on the recruitment problem a more comprehensive investigation of the recruitment processes will be necessary if this very complex system is to be understood. And, with the very high fishing pressure which most of the demersal stocks are subjected to on Georges Bank, it is very likely that it will be the effect of overfishing rather than of environmental or climatic variations that will be the more important. Thus, in the "Report of the Status of the Fishery Resources of the Northeastern United States 1989" it is concluded "that there is little evidence of longtime climatic changes that might have affected the recruitment of several species simultaneously as would be necessary to cause declines in abundance of the magnitude apparent in this data".

In referring to "this data" the report alludes to the overall conclusion that the heavy fishing pressure on commercially high-priced fish stocks is changing not only these populations but also the entire ecosystem. This not only concerns the imbalance in exploitation of the groundfish species but also the rapidly increasing and under-exploited pelagic populations.

The North Sea fisheries

Probably the most remarkable development in the North Sea fisheries during the 1980s was the increased catches of North Sea herring, as shown in Figure 22. Following the collapse of this great herring stock during the 1970s and the subsequent fishing ban from 1977 to 1980 catches have increased from 60 000 tonnes at the beginning of the decade to about 700 000 tonnes in 1988 and 1989. During the period 1948–1962 the catches were at about this level, fluctuating between 600 000 and 800 000 tonnes. However, during the middle of the 1960s and because of the collapse of the Atlanto-Scandian herring a very efficient purse seine fleet entered the northern North Sea and the catches rose very rapidly to almost 1 200 000 tonnes in 1965. This greatly increased efficiency of the fishing fleet brought about a very sharp increase in the fishing mortality from about F = 0.4 to at least F = 1 or more (Fig. 23). During the 1970s it was generally accepted, at least by the fisheries biologists working on this herring stock, that it was the increased fishing pressure which brought about the collapse of the stock and that the North Sea herring had been recruitment-overfished. Thus, during the 1970s the spawning stock was between 50 000 and 250 000 tonnes and produced very poor year classes (Fig. 24). However, at the beginning of the 1980s, when the spawning stock was still at a low level, it nevertheless started to produce strong year classes. It was this change in the level of recruitment to the stock which brought about its rapid recovery during the 1980s. Consequently, scientists started to speculate on whether the reduced recruitment of the North Sea herring during the 1970s could have been due to environmental factors. Thus Corten (1986) pointed out that during the 1970s the distribution of herring larvae in the North Sea was abnormal. Before and after the 1970s the herring larvae drifted from the spawning grounds along the east coast of Great Britain to the nursery areas in the German Bight and Jutland. During the 1970s, however, the larvae were admittedly very few but they were distributed in the centre of the North Sea and therefore never reached their normal nursery areas. Corten (1986) suggested that this was due to the reduced inflow of Atlantic water into the North Sea during the 1970s. This has been partly supported by Kamp (1990), whose calculations of surface areas between isohalines in the North Sea from 1970 to 1989 concluded that there was a weak inflow of Atlantic water

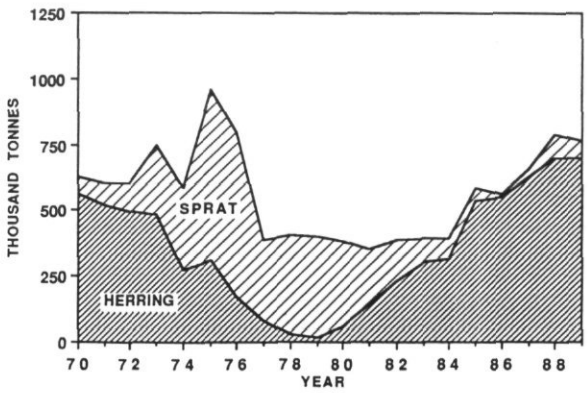

Figure 22. Annual North Sea herring and sprat catches, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

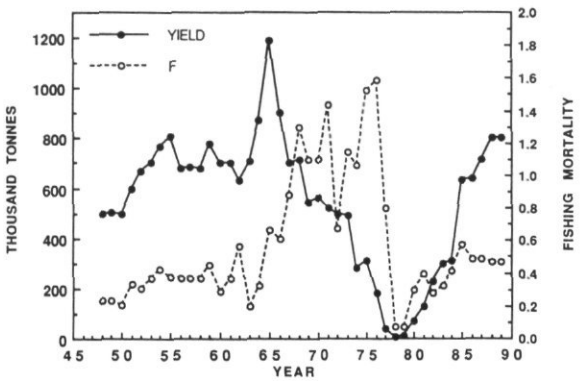

Figure 23. Yield and fishing mortality in North Sea herring 1948–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

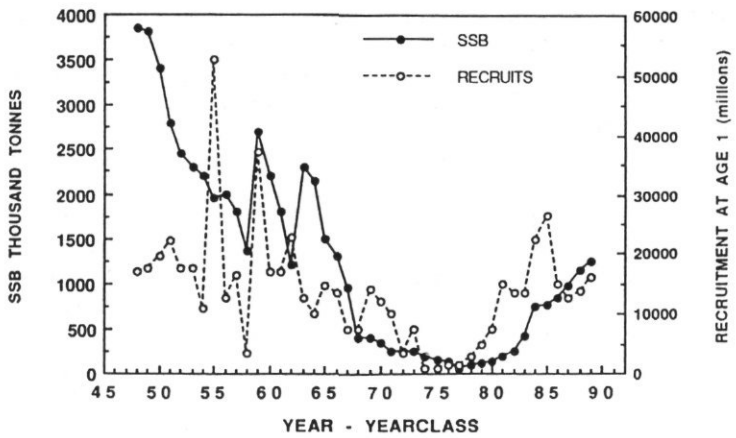

Figure 24. Spawning-stock biomass and recruitment in North Sea herring, 1948–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

during the periods 1977–1981 and 1987–1989. The "Great Salinity Anomaly" was only present in the North Sea from 1976 to 1979 while there had been a decreasing trend in the recruitment since 1969. However, the real recruitment failure occurred during the period 1974 to about 1978. It has also been pointed out that there was a sharp reduction in the populations of small copepods in the North Sea during the 1970s and a reversal at the beginning of the 1980s (Colebrook *et al.*, 1984). This, too, may have affected the recruitment levels of the North Sea herring, because these copepods are the most common food for herring larvae.

The small spawning stock of herring did not produce year classes of normal abundance in the 1970s while a similar small spawning stock did at the beginning of the 1980s. This, however, says very little about how a large spawning stock would have fared during the 1970s in the North Sea; larval production would have been at a much higher level and their distribution much wider, both in time and space. Thus Munk and Cristensen (1990) maintain that the scarcity of herring larvae in the southern parts of the North Sea described by Corten (1986) for the poor 1976–1978 year classes can likewise be explained by poor spawning off Buchan and/or Yorkshire. But these spawning grounds (especially off Buchan) were almost abandoned during the 1970s. The collapse of the North Sea herring during the 1970s is yet another example illustrating that there is usually more than one reason for such an event.

During the 1980s the catches of the four main gadoid species – cod, haddock, whiting, and saithe – in the North Sea decreased from about 800 000 tonnes in 1980 to just over 400 000 tonnes at the end of the decade (Fig. 25). The catches of cod were about 250 000 tonnes at the beginning of the 1980s but have now been reduced to about 120 000 tonnes; similarly, the catches of haddock decreased rapidly during the last three years of the decade. These two species have been exploited at a very

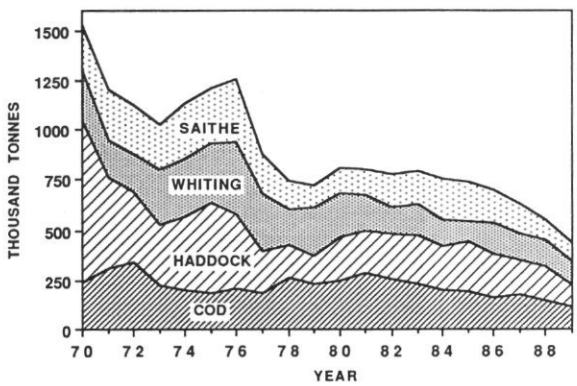

Figure 25. Annual North Sea roundfish catches, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

high rate (Figs. 26 and 27), and, as a result, the present cod catches mainly consist of one and especially two-year-old fish. This means that each year the catches are very much dependent on the recruiting year class. With a decreasing trend in recruitment since 1980 (Fig. 28) it is not surprising that there is a similar trend in the cod catches. With this fishing pattern it can safely be said that the North Sea cod has been growth-overfished for quite a long time. Similarly, there has been a decreasing trend in the recruitment to the North Sea haddock stock in recent years (Fig. 29). During the 1980s only two reasonable year classes and eight poor ones have been produced. On the other hand, there has been an increasing trend in the recruitment of whiting during the 1980s. Whiting has been exploited at a somewhat lower level of fishing mortality than the North Sea cod and haddock.

The catch figures given in Figure 25 are taken from the Roundfish Working Group reports, as presented in the ACFM reports of ICES. ACFM has, however, repeatedly pointed out that the nominal catch statistics for

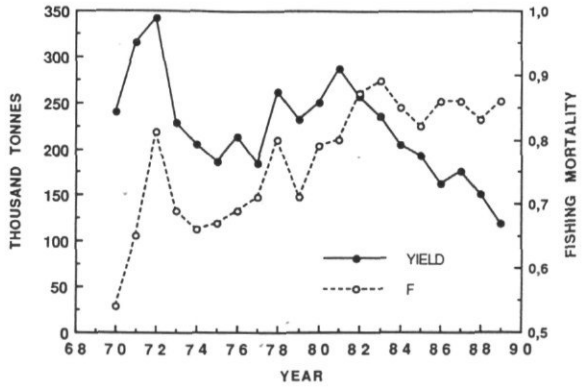

Figure 26. Yield and fishing mortality in North Sea cod, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

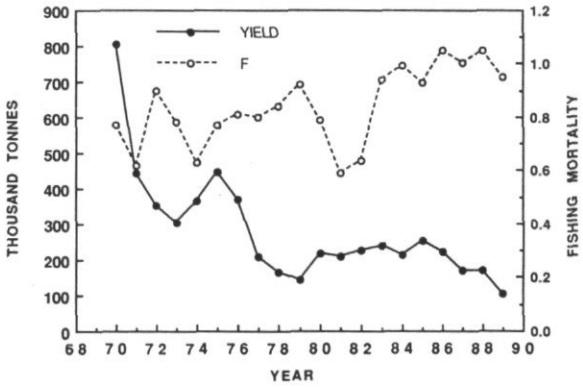

Figure 27. Yield and fishing mortality in North Sea haddock, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

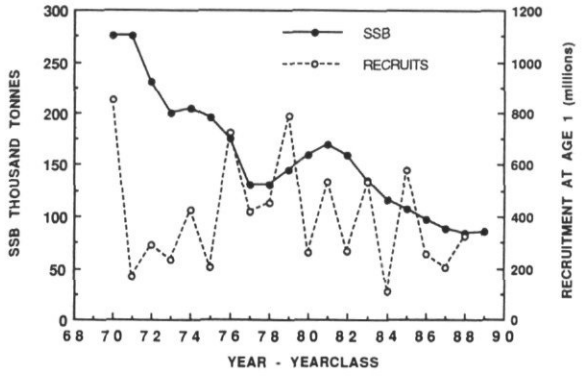

Figure 28. Spawning-stock biomass and recruitment in North Sea cod, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

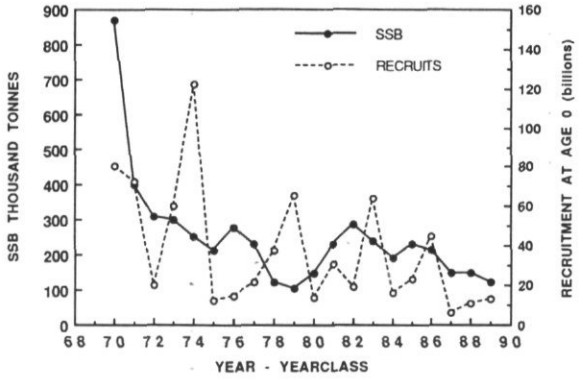

Figure 29. Spawning-stock biomass and recruitment in North Sea haddock, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

these species have become increasingly faulty and unreliable. So, actual catches may be somewhat different from those presented here. However, the fact remains that the exploitation rate of cod and haddock in the North Sea has been extremely high in recent years, with a corresponding decrease in stock abundance.

During the 1980s flatfish catches in the North Sea have continued at a very high level. Similarly, industrial fish landings have remained very high, although there has been a shift from Norway pout to sand eel as the major industrial species (Fig. 30).

In this connection it is worth noting that the rise in abundance of the gadoid stocks in the 1960s occurred during the period of cooling (Cushing, 1982) and that this was probably associated with the delay in the seasonal production of *Calanus* on which the gadoid larvae depend. However, Brander *et al.* (1992) were unable to confirm this when analysing Continuous Plankton Recorder data for the area concerned.

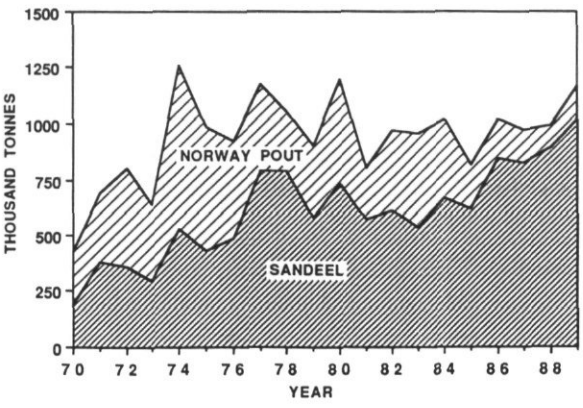

Figure 30. Annual North Sea industrial landings of sandeel and Norway pout, 1970–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

The increased yield of flatfish stocks, especially sole, during the 1930s and the 1950s has been associated with the general warming of the area. These stocks do not seem to have been affected by the cooling during the 1960s and the 1970s. The very high rate of exploitation at present makes it very difficult to detect the effect of environmental changes. It would, however, be an interesting (although unlikely) experiment to arrange a five-year cessation of fishing to see if this would have the same effect in the 1990s as in the 1940s – i.e. 50 years ago.

The fisheries in the Baltic

The fisheries in the Baltic are primarily based on three species, i.e. cod, herring, and sprat. In addition, there are fisheries for salmon and flatfish but on a much smaller scale. During the late 1970s and the first half of the 1980s the yield of the cod fisheries was at a higher level than had been experienced for several decades, reaching a maximum of about 450 000 tonnes in 1984 (Fig. 31). From 1985 until 1989 there was a steady decrease in yield in the cod fisheries with the catch in 1989 being less than 200 000 tonnes. This is at about the same level as the yield of the cod fishery in the early 1960s. The reason for the decline in the cod fisheries towards the end of the 1980s is considered to have been due mainly to the reduced recruitment to the cod stock but at the same time to the exceptionally high fishing mortality, especially during the last two or three years.

The yield from the herring fishery in the Baltic has been exceedingly steady during the last 20 years, varying between 370 000 and 450 000 tonnes (Fig. 32). Throughout this period the fishing mortality has been at a low level in all herring stocks except that in the Southwestern Baltic. The yield of the sprat fishery was at a low level at the beginning of the 1980s but increased to some extent towards the end of the decade (Fig. 32). This has been associated with decreased predation by the cod stock.

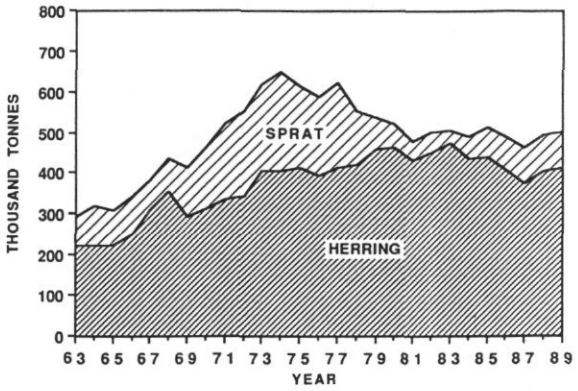

Figure 32. Annual landings of Baltic herring and sprat, 1963–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

The Barents Sea

The yield of the fisheries in the Barents Sea during the 1980s was at a low level for all the demersal species. The landings of cod varied from 300 000 to 500 000 tonnes with an average of about 370 000 tonnes compared to average landings of about 780 000 tonnes during the 1970s (Fig. 33). Fishing mortality was high until 1989 (Fig. 34). Similarly, landings of the Northeast Arctic haddock ranged from 17 000 to 150 000 tonnes, with an average of less than 70 000 tonnes during the 1980s compared to about 160 000 tonnes during the 1970s. Yields of other stocks like Northeast Arctic saithe, redfish, and Greenland halibut were also at a very low level during the 1980s.

The Northeast Arctic cod is potentially the largest cod stock in the world, with annual catches historically reaching about one million tonnes. The very low catches in the early 1980s were due to a series of small year classes during the cold period from 1976 to 1982 (Fig.

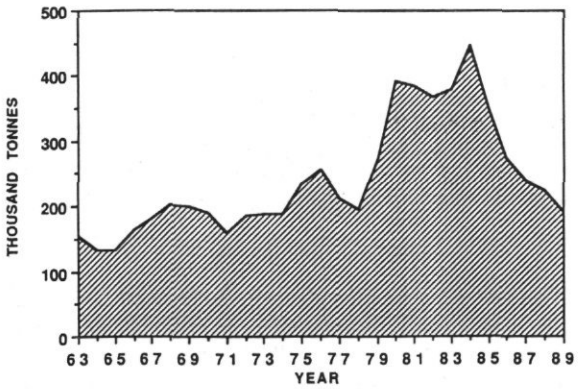

Figure 31. Annual landings of Baltic cod, 1963–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

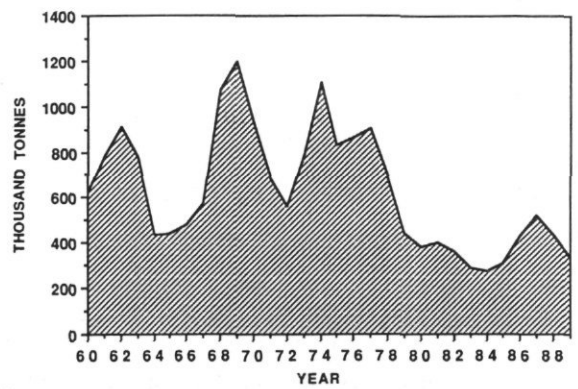

Figure 33. Annual landings of Northeast Arctic cod, 1960–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

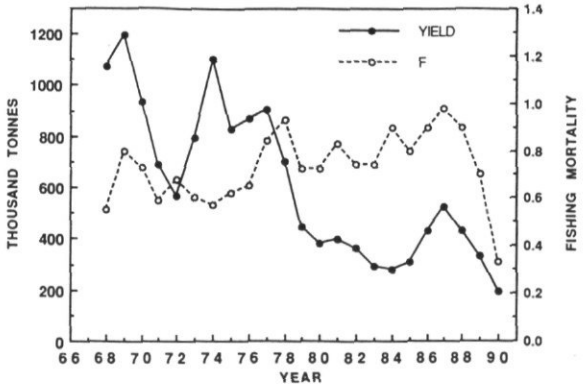

Figure 34. Yield and fishing mortality in Northeast Arctic cod, 1968–1990. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

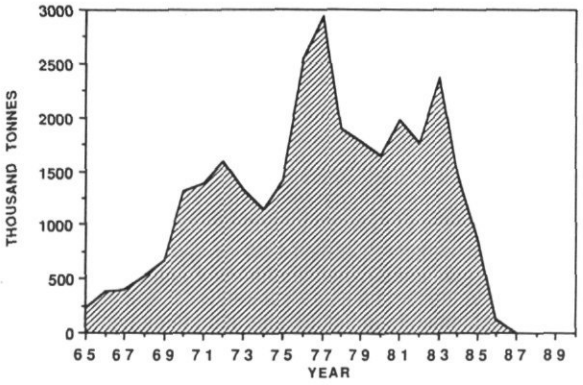

Figure 36. Annual catches of Barents Sea capelin, 1965–1990. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

35). Dickson *et al.* (1988) showed that the "Great Salinity Anomaly" peaked on the west coast of Norway in 1977–1978 and in the Barents Sea in 1978–1979. Associated with a warmer marine climate in 1983, a very strong year class was produced and was followed by two year classes (1984 and 1985) which originally were classified as above average. As a result, scientists predicted rapid increases in both stock abundance and catches towards the end of the 1980s and hopes were high in the middle 1980s within the fishing industry and among the scientists.

During the period 1980–1984 catches of the Barents Sea capelin ranged from about one-and-a-half million to almost two-and-a-half million tonnes (Fig. 36). There was a decline in the catches to 870 000 tonnes in 1985 and a further decline to only 123 000 tonnes in 1986, whereupon the fishery was closed.

The collapse of the capelin stock in 1986 had disastrous effects on the ecosystem. Traditionally, capelin and juvenile herring have been the most important food species for the cod (age 3+) in the Barents Sea. With the herring stock still at a low level after the collapse in the late 1960s the capelin stock became all the more important; in the mid-1980s the rapidly growing cod stock grazed down other available food, including its own progeny (Mehl, 1989). From 1986 onwards the cod suffered starvation and the average weight-at-age dropped by about 50%. The invasion of Arctic harp seals into Norwegian coastal waters in search of food accentuated the disaster and led to a degradation of the Barents Sea ecosystem. This was all the more dramatic because of the high expectations of increased fisheries from the middle 1980s. However, because of the great reduction in the fish-eating stocks, i.e. primarily the cod stock, the predator–prey balance was re-established by the beginning of the 1990s and the ecosystem in the Barents Sea seems to be recovering.

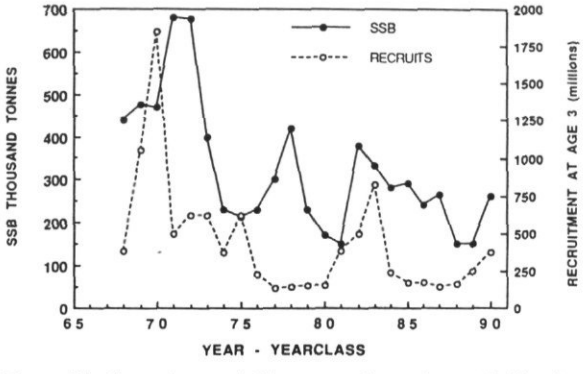

Figure 35. Spawning-stock biomass and recruitment in Northeast Arctic cod, 1968–1990. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

There has been much discussion concerning whether these dramatic events in the Barents Sea during the late 1980s were due to natural or man-made causes. Some scientists maintain that the fisheries in the 1980s had very little to do with the ecological collapse (Hamre, 1990) and that its sources are to be found in the collapse of the Atlanto-Scandian herring in the 1960s. Hopkins and Nielsen (1992), on the other hand, consider that the present evidence clearly implicates the capelin fishery as an important contributory factor to the lost potential biomass of the 1983 year class of cod and the lack of food for the higher trophic levels. In hindsight, it would have been a very interesting experiment to have stopped the capelin fishery in 1983 with the first indications of strong cod, haddock, and saithe year classes, thus making about four million tonnes of extra food available for the increasing predator stocks. Whether this would have prevented the collapse of the ecosystem will never be known for certain, but it would have been worthwhile trying. It should, however, be made clear that nobody suggested this at the time.

Hamre (1990) has described the Northeast Atlantic region as one of high productivity but of low biological

diversity, i.e. it is rich in fish production but the fish belong to a few species. Such an ecosystem is attractive to fisheries, but they have low stability and may be seriously affected by over-exploitation of one or more key species in the system. In this case there were two key species in the system, the herring and the capelin, which have served as food for the predators of the Barents Sea. Once the herring had been depleted hindsight tells us that greater care should have been taken in conserving the other key species (Hopkins and Nielsen, 1992). This was not done and the capelin was fished until the stock collapsed. In order to illustrate the magnitude of the Barents Sea crisis during the late 1980s, the estimates of the 1983–1985 year classes of cod, as assessed in 1986 and 1990, are given in the text table below. It should be noted that the 1986 assessments of the 1984 and 1985 year classes were preliminary and based primarily on 0- and 1-group surveys.

Number of recruits (millions) as three-year-olds

	Year classes			
Assessment year	1983	1984	1985	Total
1986	1500	800	1100	3400
1990	790	235	174	1200
Difference	710	565	926	2200

In 1986 recruitment from these three year classes was expected to be 3400 million three-year-old cod, whereas in the 1990s the estimates were only 1200 million three-year-old recruits. The difference, 2200 million, corresponds to a difference in yield of about three million tonnes from these three year classes alone.

Although the 1980s have been a very difficult period for fisheries scientists working in the Barents Sea great strides forward have been made in our understanding of the importance of climatic variations in this area. Just a few examples will be mentioned here. Ellertsen *et al.* (1989) have described the relation between temperature and the survival of eggs and first feeding larvae of Northeast Arctic cod. They found that only small year classes are produced when temperatures are low in the distribution areas of the early larval stages, while year classes of all strengths are produced under higher temperatures (Fig. 37). Thus, in 1983, unusually high temperatures (3.6°C) were observed in the coastal watermasses during spawning. This was the highest temperature in this area since 1961. In 1981, only two years earlier, eggs and the early larvae experienced the lowest environmental temperature ever recorded since temperature measurements began in 1936. The relationship between year-class strength at age 3 and the temperature during the egg/early larval stages indicates that elevated temperature is positively linked with the formation of strong year classes. It also implies that other processes are important, since strong year classes are not always produced in all years with high temperatures. The significance of temperature is also supported by the work of Sætersdal and Loeng (1987), who found that strong and medium year classes occurred most frequently in years with high temperatures along the Kola section in the period 1900–1983. Since most of the temperature variations in the Norwegian Sea/Barents Sea are of a geographically large scale the same variations also occur in the Lofoten spawning area. The influence of temperature on the spawning success of the Northeast Arctic cod is explained by the fact that the spawning of *Calanus finmarchicus* appears to be temperature triggered (Fig. 38), while spawning of the cod seems to be more or less constant or unrelated to

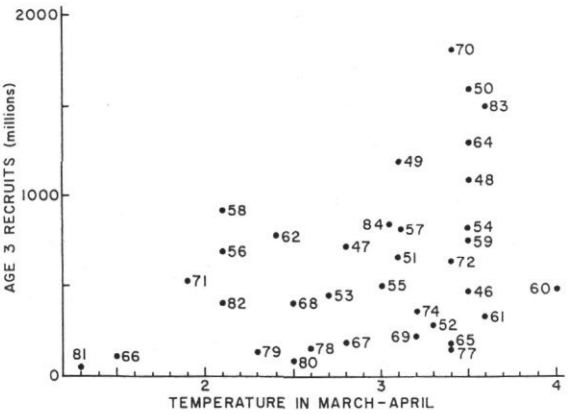

Figure 37. Year-class strength in Northeast Arctic cod and temperature on the spawning grounds. Redrawn from Loeng (1990).

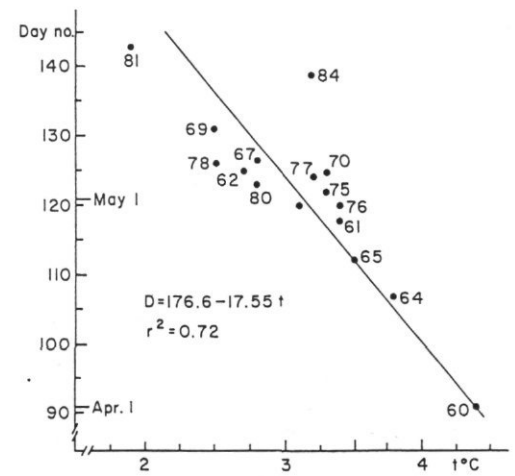

Figure 38. Time of maximum occurrence of *Calanus finmarchicus*, stage I, versus temperature. Redrawn from Ellertsen and Fossum (1989).

temperature variations. In years with normal environmental temperatures a high proportion of cod larvae will experience food concentrations high enough for growth and survival in most of the distribution area due to the match between larvae and their prey. In years with extreme temperatures, both high and low, the mismatch in time may contribute significantly to the reduced potential of year classes produced in these years.

The year-class strength is affected not just by the temperature but also by the growth rate at later stages. Loeng (1990) has shown (Fig. 39) that three-year-old cod in the western Barents Sea grows at almost twice the rate of cod of the same age in the extreme eastern part of the Barents Sea and that there is a clear gradual reduction in the growth rate from west to east. Similarly, the distribution of fish larvae is greatly affected by the strength of the water currents and temperature regime in the Barents Sea. Thus, it is quite clear that large-scale oceanic or climatic variability greatly affects marine resources in the Barents Sea. In addition, research on the feeding conditions of Arcto-Norwegian cod larvae corroborates the Rothschild-Osborne theory of small-scale turbulence and plankton contact rates (Sundby and Fossum, 1990).

The fisheries at Iceland

The most dramatic event in the Icelandic fisheries during the 1980s was probably the collapse of the capelin fishery at the beginning of the decade and the subsequent rapid recovery in 1984 and 1985 (Fig. 40). Associated with this was the great decrease in the growth rate of the Icelandic cod and the weight-at-age reduction for three- to six-year-old cod of about 25%. With the recovery of the capelin stock the weight of the cod returned to normal within two years. The collapse of the capelin stock was

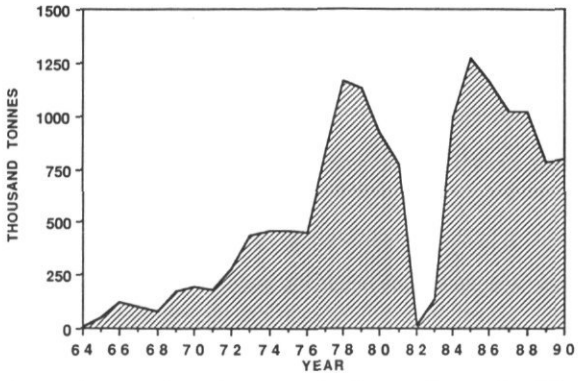

Figure 40. Annual catches of Icelandic capelin, 1964–1990. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173.

due to very poor recruitment during the period 1978–1981. By 1981 and 1982 the recruitment level increased somewhat, but it was the very abundant 1983 year class which brought the stock back to its former abundance (Fig. 41).

There is a remarkable relationship between recruitment of age 2 individuals in the Icelandic capelin and zooplankton abundance north of Iceland at the time of recruitment (Fig. 42), but this is not observed when comparing measurements of 1-group capelin and zooplankton abundance.

The cod fishery at Iceland was at a very high level in 1980 and 1981 (Fig. 43), due mainly to the immigration of the 1973 year class from Greenland to the Icelandic spawning grounds. There was a sharp decrease in the catches in 1982 and 1983 because of rapidly decreasing weight-at-age during this period, as mentioned above. During the late 1980s the cod catches at Iceland were between 300 000 and 400 000 tonnes per year and the

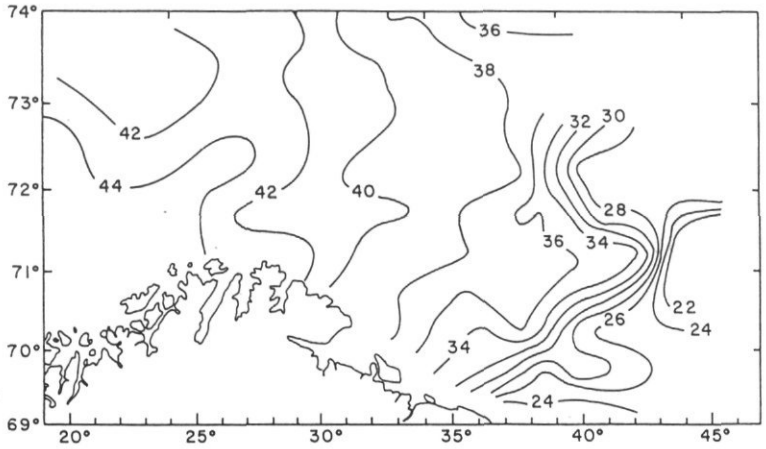

Figure 39. Length distribution of age 3 cod in the Barents Sea, February 1984. Redrawn from Loeng (1990).

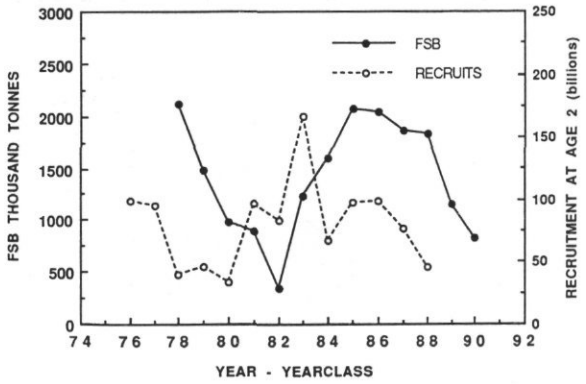

Figure 41. Fishable stock abundance on 1 August (at the beginning of the season) and recruitment of the Icelandic capelin, 1976–1990 (Hjálmar Vilhjálmsson, pers. comm.).

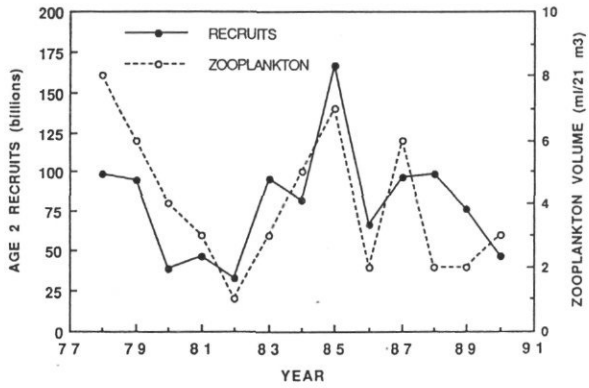

Figure 42. Relationship between the abundance of age 2 recruits in the Icelandic capelin stock and the abundance of zooplankton north of Iceland in spring of the year of recruitment (Hjálmar Vilhjálmsson, pers. comm.).

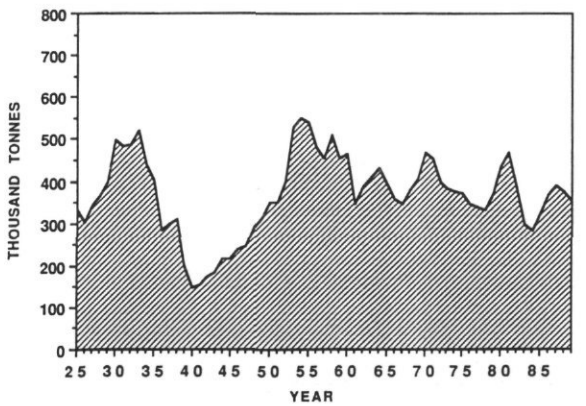

Figure 43. Annual catches of Icelandic cod, 1925–1989. Data from the Mar. Res. Inst. Reykjavík.

fishery was controlled by quota regulation. The catches of demersal stocks like redfish, haddock, and saithe were more or less stable during the 1980s.

The herring fishery, which has only exploited the Icelandic summer spawning herring since the 1960s, has gradually increased from about 50 000 tonnes at the beginning of the 1980s to 100 000 tonnes at the end of the decade (Fig. 44). The herring fishery has been strictly controlled by quota regulations and the aim has been to keep the exploitation rate at the low level of c. $F_{0.1}$ (Fig. 45). This has in fact been one of the two stocks in the northern Atlantic, outside the Baltic, where this aim has been more or less fulfilled. Results have been outstanding (Fig. 46), with the present stock abundance at a highest historical level.

No such policy has been accepted for the exploitation of demersal species in the Icelandic area, nor is it likely that fishing cod at such a low F level is desirable. The exploitation rates of saithe and haddock, however, were at a reasonable level (F = 0.4–0.6), while that of the most important species, the cod, remained at a high level (F about 0.8) during the 1980s, despite the quota regulation (Fig. 47). This was due partly to the fact that the quota system has been going through various stages of development since 1984, when it was introduced, and partly because fishery managers have not been able to withstand the temptation of trading long-term goals for short-term gains. As a result of this high exploitation rate of the Icelandic cod, the spawning stock biomass was at a low level during the 1980s (Fig. 48). The main characteristic of the recruitment to the Icelandic cod stock since the early 1950s is its remarkable steadiness. The difference between strong and weak year classes has been by a factor of only about 3 to 4, which is much less than that observed in most other cod stocks (Fig. 48). During the 1980s, recruitment to the Icelandic cod stock was at a low level, however, with one average year class and only two strong year classes; seven were below average. This has been associated with Arctic conditions

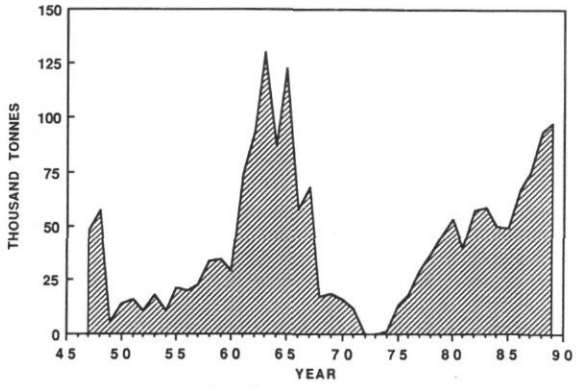

Figure 44. Annual catches of Icelandic summer spawning herring, 1947–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173 and Jakobsson (1980a).

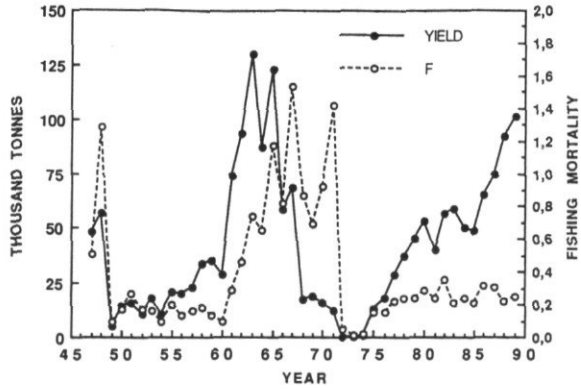

Figure 45. Yield and fishing mortality in Icelandic summer spawning herring, 1947–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173 and Jakobsson (1980a).

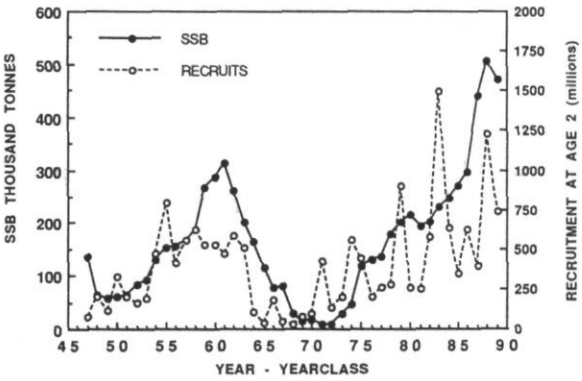

Figure 46. Spawning-stock biomass and recruitment in Icelandic summer spawning herring 1947–1989. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173 and Jakobsson (1980a).

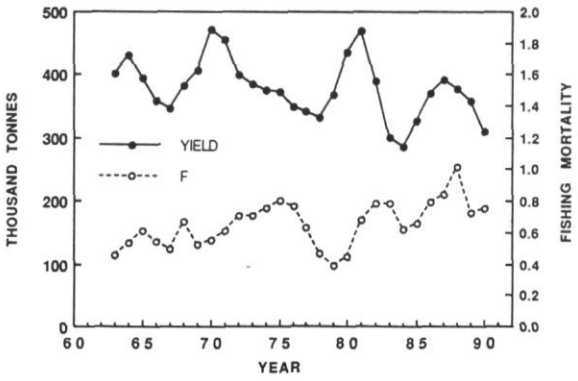

Figure 47. Yield and fishing mortality in the Icelandic cod stock, 1963–1990. Data from the Mar. Res. Inst. Reykjavík.

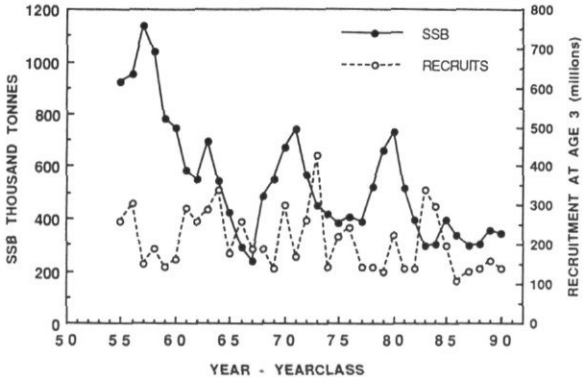

Figure 48. Spawning-stock biomass and recruitment in the Icelandic cod stock, 1954–1990. Data from the Mar. Res. Inst. Reykjavík.

in Icelandic waters at the beginning and end of the 1980s, with a warmer period in the middle of this decade. In 1983, Arctic conditions still prevailed, while in 1984 a shift to a warmer period took place. Strong year classes were produced in both years. The temperature on the spawning grounds at southwest Iceland varies very little (<2°C) from one year to the next because the southwest coast is bathed in warm Atlantic water.

The main environmental variable on the spawning grounds of cod at Iceland, one which could affect the survival or success of the cod larvae, is considered to be variation in the stratification of the surface layers. Stratification triggers the spring phytoplankton bloom, which is followed by the outburst of *Calanus* nauplii which in turn is the main food of the cod larvae. The larvae drift in a clockwise direction through Icelandic coastal waters during their first months of life and thus enter an area with much greater hydrographic variability, off the north and east coasts.

During the 1980s the two good year classes of cod came at the end of Arctic conditions and at the beginning of a warm period. This suggests that the size of the year class is not finally determined until after the first winter. Although the 1983 year class was spawned during Arctic conditions off north Iceland these had no effect on the spawning grounds, and by the time the larvae had drifted to the area north of Iceland the warming had started in the late autumn of 1983. This is very similar to what often happens in the Northeast Arctic cod stock. Large year classes are produced either during a warm period or just before its beginning.

Strong 1983 year classes

One of the interesting events during the 1980s was the production of a strong 1983 year class in several fish stocks in Icelandic and Norwegian waters. At Iceland the species concerned were cod, saithe, herring, and capelin. The 1983 year classes in the Icelandic summer

spawning herring stock and the Icelandic capelin stock were the largest on record. In Norwegian waters the species concerned were cod, saithe, and haddock in the gadoid group, and the 1983 class of herring has been by far the largest during the last 20 years. Thus, out of the 10 stocks considered, 8 produced strong year classes in 1983 (Figs. 49 and 50). Only the Icelandic haddock and the Barents Sea capelin failed to do so. The change in recruitment level of so many major fish stocks in such a wide area could be associated with the increase in inflow of warm Atlantic water to the Icelandic and Barents Sea areas. At Iceland the 1983 year class was followed by strong cod and saithe year classes in 1984. Both the 1984 and 1985 haddock year classes were also strong. In the Barents Sea the recruitment potential was not realized owing to food limitation.

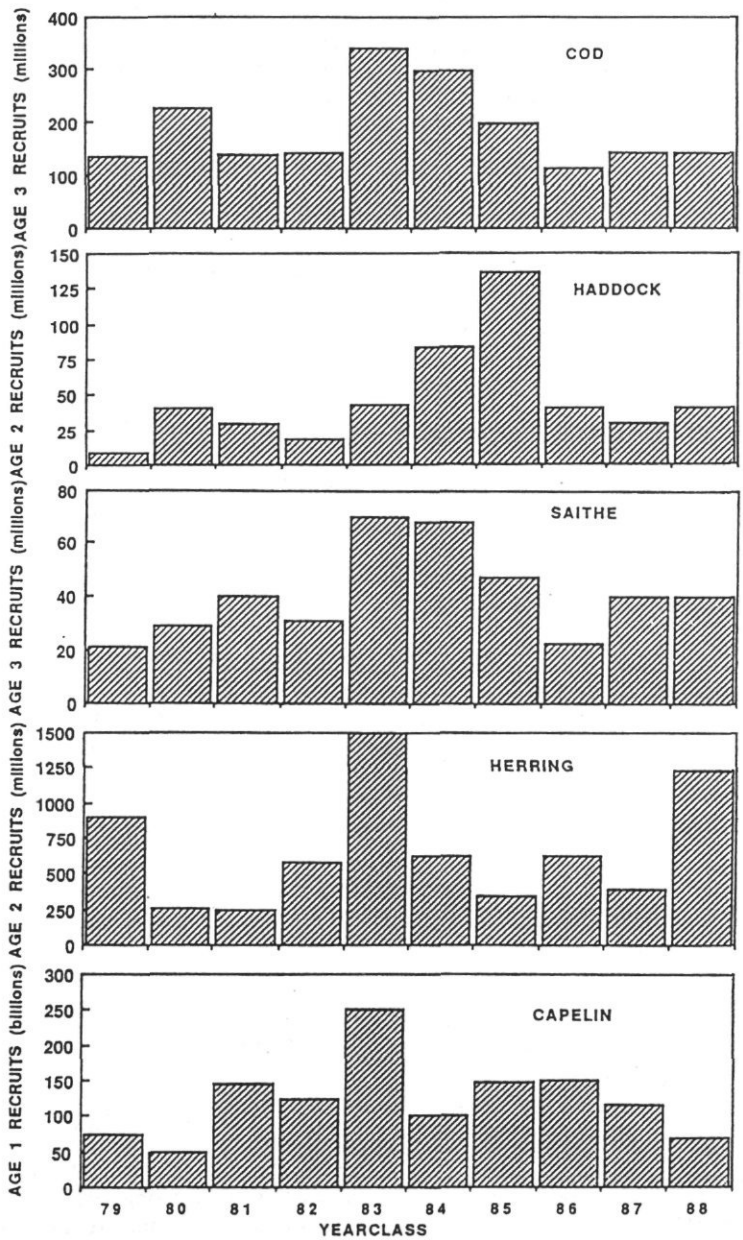

Figure 49. Recruitment of recent year classes in five Icelandic fish stocks. Data from the Mar. Res. Inst. Reykjavík.

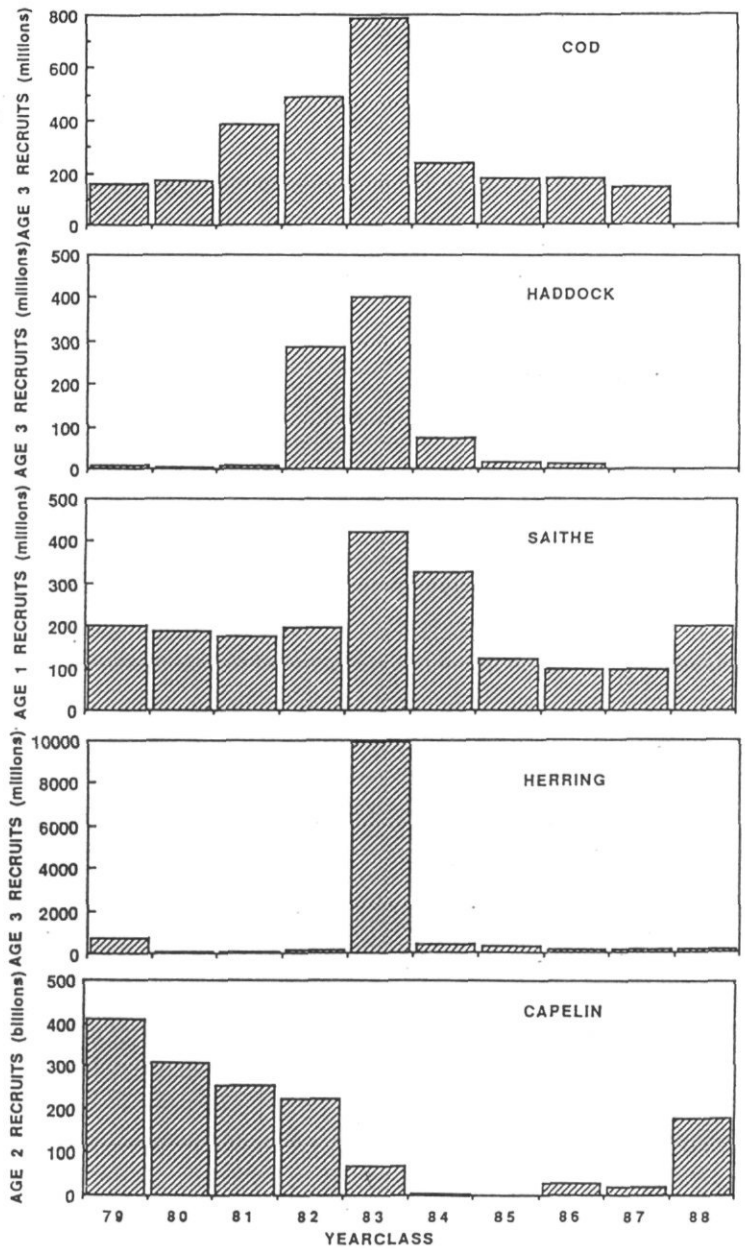

Figure 50. Recruitment of recent year classes in five Barents Sea fish stocks. Data from Coop. Res. Rep. Cons. int. Explor. Mer, 173 and associated W.G. Reports.

Discussion

The 1980s have been an eventful decade in terms of the variability in fisheries. The reasons for the great variations are almost certainly both natural and man-made. In this attempt to overview these complicated events I have deliberately followed the path of the "Great Salinity Anomaly", because its presence has coincided with greatly reduced recruitment in several important fish stocks. This in turn has caused variability in the fisheries.

Just before the beginning of the 1980s the management of the fisheries was shifting from management by international organization to management by coastal state. International management of fish resources continued however in areas outside the 200-mile limit. In some areas two or more coastal states had to negotiate the management of the fish resources within their respective 200-mile zones.

In the two southernmost areas discussed, i.e. off the northeastern coast of the United States and in the North

Sea, no effective fishery management was enforced. In both areas the highly priced demersal resources were exploited at a very high rate of fishing. Under these circumstances it is often very difficult, if not impossible, to detect variations due to environmental fluctuations. It is generally accepted that in both these areas practically all the demersal and highly priced species are overfished, some even recruitment-overfished, although this is very difficult to prove. At the same time there has been a dramatic recovery of the mackerel stock in the west Atlantic and the herring stock in the North Sea. The North Sea herring fishery was at about the same level as in the 1950s and early 1960s before the increased level of fishing mortality on the stock and its subsequent collapse in the 1970s. The mackerel fishery on the west side of the Atlantic, on the other hand, has not matched the increase in stock abundance and is considered underexploited. Although there are many similarities between the North Sea and the fishing areas off the northeastern United States (e.g. the lack of management of the stocks, almost uncontrolled fishing, the fact that these are the two most productive regions per unit area in the North Atlantic), there are of course many differences. One is a more or less enclosed marginal sea area, the other is not, and the physical oceanography is very different in the two areas. It is, however, difficult to visualize how the ecosystem of any other areas in the North Atlantic could withstand such a heavy rate of exploitation as has been the case in the North Sea and off the northeastern coast of the United States.

The declared management policy of Canada is the opposite of the management of fish resources in the North Sea and off the northeastern United States. Although the Canadians have not succeeded in rebuilding their demersal fish stocks during the 1980s, the fisheries in Canadian waters have been very stable during this period, fish sources, especially those which are fully under Canadian management, have been exploited at a more reasonable level than in most other areas of the Atlantic. It should also be noted that the fish resources in these waters were in a depleted state at the end of the 1970s. And if it is the case, as the data indeed indicate, that the main cod stock, the northern cod, has been recruitment-overfished, it will take a long time before the productivity of the 1960s returns.

Management of Icelandic fish resources has evolved during the 1980s. The Icelandic authorities have accepted exploitation of the local herring stock at the same low level as Canadian policy, i.e. to fish at $F_{0.1}$. This policy has proved very successful. The cod stock, however, was exploited at a high rate throughout the 1980s. Catches varied between about 470 000 tonnes at the beginning of the 1980s and less than 300 000 tonnes in the middle of this period. These fluctuations are associated with the return of the 1973 year class from Greenland in 1980 and 1981 and the collapse of the capelin stock in 1982–1983. This happened during the period when Icelandic waters were revisited by the "Great Salinity Anomaly".

Of the areas discussed in this paper the most dramatic events occurred in the Barents Sea, where retrogressive occurrences took place in the ecosystem during the latter half of the 1980s.

One lesson we have learned from events in the fisheries, in the Barents Sea and elsewhere, during the last decade has been that sound management policy for the fish resources has to be based on a thorough knowledge of all the characteristics of each ecosystem. Although there are similarities between some of these large ecosystems there are also clear differences. Thus, the ecosystem in the Baltic, i.e. in the sea around the islands where this symposium was held, is characterized by relatively few species. There is one major predator species, the cod, and two prey species, the herring and the sprat. In addition, there is a major and imminent threat from pollution in the area. By contrast, the ecosystem in the North Sea is based on a close interaction of many species living in a relatively stable physical and chemical environment.

The ecosystem in the northeastern Arctic, the Barents Sea, is fairly similar to that of the Baltic in that there are relatively few species, with cod and haddock as the main predator species and herring and capelin as the main prey species for the larger gadoids. There is, however, one very important difference, because in the Northeast Arctic the marine mammals play a very important role in the ecosystem. The Baltic, North Sea, and the Barents Sea have the common physical characteristics of shallow marginal seas with close interaction of the species living there, and the systems are very much dependent on the inflow of water from the outside. Moving further west to Iceland, one comes to an ecosystem around an island which is much more open and where interactions of species are less prominent than for instance in the North Sea. The cod is the main predator fish while capelin is the main prey species. Because of the openness of the system both the capelin and the herring undertake large migrations during certain parts of their life cycle, far beyond the continental shelf and far beyond the reach of their main predator, the cod. Again, marine mammals play a very important role in the ecosystem at Iceland.

This is also true of Canadian waters, a large and complex area with several major ecosystems, each with its own characteristics. This may also be true of the waters off the coast of the northeastern United States, where the Georges Bank ecosystem appears to be very different from that in the Gulf of Main.

Variations in the physical properties of the ocean play a very important role in all the Arctic and Subarctic ecosystems, which is somewhat contrary to the features of ecosystems at lower latitudes. Thus it is easier to detect the role of oceanic processes in Arctic and Subarctic regions.

As the time series for stock fluctuations and environ-

mental data have lengthened it has become more and more evident that major variability in fisheries usually has at least two different sources, i.e. anthropogenic sources and environmental sources. A high exploitation rate places stress on the marine populations, rendering them vulnerable to changing environmental effects. In recent years there have been several examples of this, at least some of them associated in one way or another with the "Great Salinity Anomaly", as has been discussed in this paper.

The most important lesson to be learned is that each of the large ecosystems has its own features, and a thorough knowledge of these is the basic requirement for a sound management policy that will lead to rational harvesting of the living resources without upsetting the all-important balance within each system.

Acknowledgments

I am indebted to several ICES colleagues on both sides of the Atlantic for responding positively to my requests for information. Dr R. S. Bailey and Professor C. C. E. Hopkins kindly read through the manuscript and suggested many improvements. In particular I thank Hjálmar Vilhjálmsson, who kindly produced most of the figures, read through the manuscript, and made many valuable suggestions. I am also greatly indebted to Dr O. S. Ástlhvorsson, who finalized the manuscript in my absence in hospital.

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In addition to these references most of the data presented in this paper are taken from official reports, such as the ACFM reports published in the *ICES Cooperative Research Report* series, the CAFSAC reports, and the reports on the Status of the Fisheries of the Northeastern Coast of the United States.

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ICES mar. Sci. Symp., 195: 316–325. 1992

Global greenhouse effects, multi-decadal wind trends, and potential impacts on coastal pelagic fish populations

Andrew Bakun

Bakun, A. 1992. Global greenhouse effects, multi-decadal wind trends, and potential impacts on coastal pelagic fish populations. – ICES mar. Sci. Symp., 195: 316–325.

Potential linkage of long-term trends in sea surface windstress in the North Atlantic to global climate change is investigated. Seasonal and locational aspects of the trends appear to correspond to a proposed "greenhouse" mechanism. A potential artifact related to increasing frequency of anemometer measurements is addressed, and several types of information tending to substantiate the reality of the indicated trends are presented. Potential alterations in suitability of reproductive habitats of coastal pelagic fishes are suggested.

La liaison potentielle entre les tendances à long terme de la tension du vent de surface en mer et le changement global du climat est examinée. Les aspects saisonniers et locaux des tendances semblent correspondre à un mécanisme de type "effet de serre". Un artefact potentiel dû à l'accroissement des fréquences des mesures anémométriques est abordé et plusieurs sources d'information tendent à prouver la réalité des tendances présentées. Les altérations potentielles des zones de reproduction des espèces pélagiques côtières sont suggérées.

Andrew Bakun: Pacific Fisheries Environmental Group, NOAA/NMFS, Southwest Fisheries Science Center, PO Box 831, Monterey, California 93942, USA.

Introduction

The existence of interdecadal trends both in ocean characteristics (e.g. Dickson *et al.*, 1988a) and in marine biological populations (e.g. Corton, 1990) has been well documented in the North Atlantic as well as in other regions of the world's oceans (e.g. Sharp and Csirke, 1983). For example, Dickson *et al.* (1988b) have pointed out a major long-term increase in northerly winds during spring and summer over the eastern North Atlantic and related that to important changes in biological populations in the region. Bakun (1990) showed that corresponding increases in wind have apparently also occurred in other eastern ocean boundary regions, and suggested a linkage to "global warming" (Schneider, 1990; Anon., 1990). The linkage to the progressive accumulation of anthropogenic greenhouse gases in the earth's atmosphere is of great interest, introducing a powerful degree of predictability into important phenomena which have heretofore been unpredictable. It also carries an implication of rapid progressive extension of climatological trends beyond the range of historical experience and perhaps beyond the limits accounted for in biological "life style" adaptations.

On the other hand, other investigators (Ramage, 1987; Cardone *et al.*, 1990) have pointed out that indications of long-term increases in winds over the oceans may be merely artifacts related to the monotonically increasing proportion of anemometer measurements to Beaufort estimates in the available distributions of maritime wind reports. Clearly, among the most important questions regarding long-period hydrobiological variability are: (1) Are the reported increasing trends in spring and summer winds over the coastal oceans real? (2) If real, are they linked to global climate change?

Linkage to resource dynamics – results from SARP

The Sardine-Anchovy Recruitment Project (SARP), the initial focal project of the Programme of Ocean Sciences in Relation to Living Resources (OSLR) which is co-sponsored by the Intergovernmental Oceanographic Commission and the Food and Agriculture Organization of the United Nations, has now been in existence for over six years (Anon., 1983; Anon., 1989). The small, pelagic-spawning, clupeoid fishes that are the focus of SARP comprise a major portion of the world's fisheries production. In addition, they constitute an

extremely important food base for other valuable fish species in many of the large marine ecosystems of the world. The SARP studies have tended to implicate three general types of factors which act in combination to produce favorable reproductive habitat: (1) enrichment of the food web via physical processes (upwelling, mixing, etc.); (2) opportunity for concentrated patch structure of food particles to accumulate (i.e. stability, lack of active turbulent mixing, and/or strong convergence in frontal structures); (3) availability of mechanisms promoting retention of reproductive products within the neritic habitat. All of these are substantially linked to wind intensity and pattern, but the relationships differ (in fact, in the case of factor types (1) and (2), the functional relationships to wind intensity tend to be in opposition). Accordingly, the three types of factors are mutually supportive only within special environmental configurations (Bakun, in press). Thus, greenhouse-related wind changes would seem capable of rearranging the geography of suitable spawning habitat, irrespective of any local temperature changes caused by direct greenhouse warming.

Mechanism for wind intensification

Bakun (1990) described a mechanism by which greenhouse-related amplification of heating during spring and summer intensifies the thermal low pressure cell over the coastal landmass. This leads to an enhanced onshore–offshore pressure gradient between this continental thermal "low" and the oceanic "high" offshore. The stronger pressure gradient induces strengthened alongshore geostrophic wind, causing enhanced offshore Ekman transport and amplified coastal upwelling on either side of a continent and in either the northern or southern hemisphere (Fig. 1). The increased upwelling and mixing implied by such a wind increase, by cooling the coastal ocean surface layers and further accentuating the ocean–continent temperature contrast, might constitute a positive feedback loop (Bakun, 1990) leading to further amplification of the effect.

Eastern sides of oceans are characterized by a much drier atmosphere than western sides. Since the most important greenhouse gas in the atmosphere is water vapor, the global accumulation of CO_2, methane, chlorofluorocarbons, etc., will have a larger relative "greenhouse effect" along eastern ocean boundaries than in other ocean regions. Note that the Atlantic ocean–atmosphere system is much more subject to continental influence than is the Pacific system. Thus large-scale effects of such a process might be expected to be more immediately evident in the Atlantic.

Evidence from other regions

Bakun (1990) presented indications of substantial multi-decadal increases in upwelling-producing wind stress off California, off Peru, off Morocco, and off the Iberian Peninsula. Roy (1990) indicated similar trends along much of the Northwest African Coast. Although the data are much more sparse, indications of corresponding increases also appear in the maritime report files from other important eastern ocean upwelling systems (e.g. off Chile and off South Africa, Fig. 2).

The multi-decadal upward trend is pronounced in the various series presented by Bakun (1990); however, a widespread tendency for the most rapid increases to be from the mid-1950s to the mid-1970s is evident. After this period the trends in widely separated regions, in both the northern and southern hemispheres and in both the Atlantic and Pacific, appear to level off, or even to "flip over" to negative trends (see also Roy (1990)).

An explanation of this "flip over" feature is not available at present. However, the Pacific system is known to have been in a state of somewhat elevated "El Niño" character since the mid-1970s. The waters of the Pacific Ocean constitute the greatest mass of the heat storage capacity on earth. Its atmosphere is only minimally subject to continental effects, and so is not nearly so strongly forced into seasonal climatic regularity as are the atmospheres of more continentally influenced oceans. The result is that the coupled ocean–atmosphere system of the Pacific "wanders" from year to year, in response to its own internal dynamics, to a much greater degree than do other regions of the globe. The most dramatic expression of this interyear variability of the Pacific system is the El Niño–Southern Oscillation (ENSO) phenomenon. Because the Pacific is so large, its effect is global. Accordingly, ENSO constitutes the dominant mode of short-term climatic variability throughout the world. Note that Inoue and O'Brien (1987) have reported an apparent decrease in strength of the trade winds in the tropical Pacific during the late 1970s. It is reasonable to speculate that the interruption of the upward trends in the coastal upwelling systems in the mid-1970s may be a global response to internal variability in the Pacific ocean–atmosphere system, which may be superimposed on the proposed underlying longer-term greenhouse-related global increase.

Spring–summer trends in wind in the North Atlantic

In order to examine the situation in the North Atlantic, time series of monthly resultant windstress estimates were produced for the period at a number of locations (A through O, Fig. 3) according to the methodology of Bakun (1973). Components were resolved along the dominant directions (Fig. 3) of wind flow around the Subtropical Gyre during spring and summer, and averaged for the months of April through September for each year since 1946 (Fig. 4). Tests for linear trends were

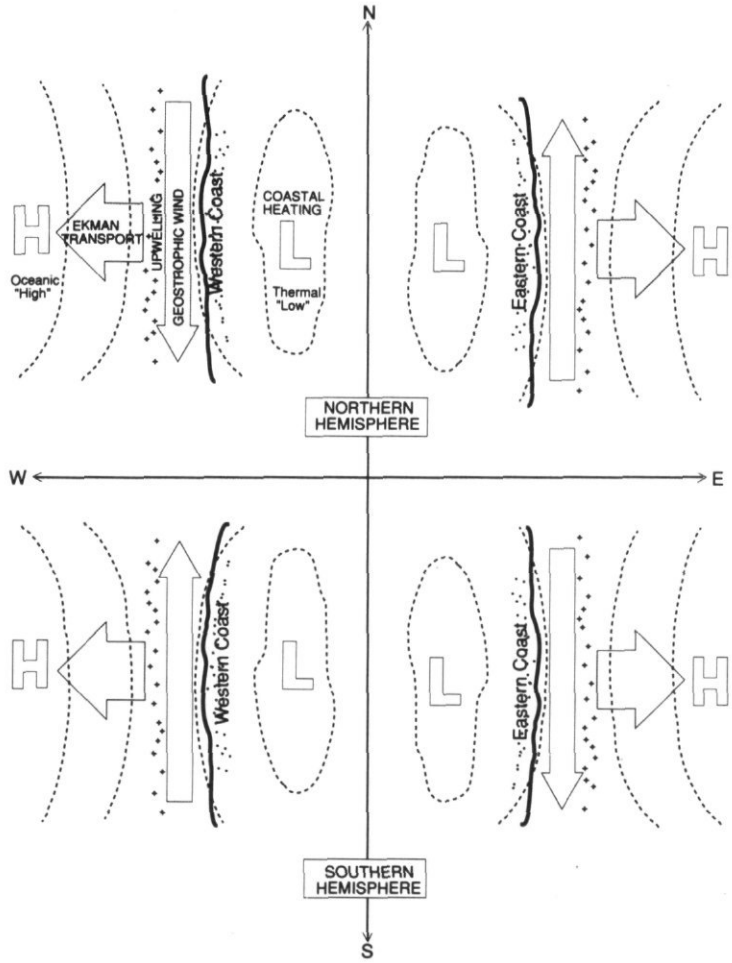

Figure 1. Proposed mechanism. Increased seasonal heating due to atmospheric greenhouse effect increases temperature contrast between continental land mass and coastal ocean. The thermal low pressure cell in the continental interior is intensified, increasing the cross-shore pressure gradient between the spring–summer continental "Low" and the offshore oceanic "High". This increased pressure gradient supports an intensified alongshore geostrophic wind directed such that, with respect to a person facing downwind, low pressure would be on his left in the northern hemisphere and on his right in the southern hemisphere. Correspondingly increased Ekman transport, directed 90° to the right of the wind in the northern hemisphere (to the left in the southern hemisphere), leads to enhanced coastal upwelling which may feed back as further enhanced land–sea temperature contrast.

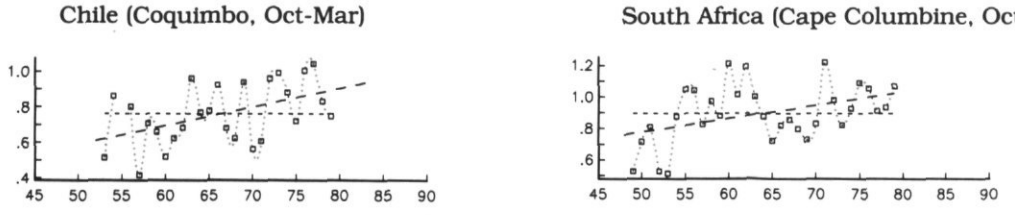

Figure 2. Spring–summer (Oct–Mar) averages of monthly windstress estimates produced from areal summaries of maritime wind reports. (a) Chile (30°S to 37°S, from the coast to 75°W). (b) South Africa (32°S to 35°S, 16° to 18°E).

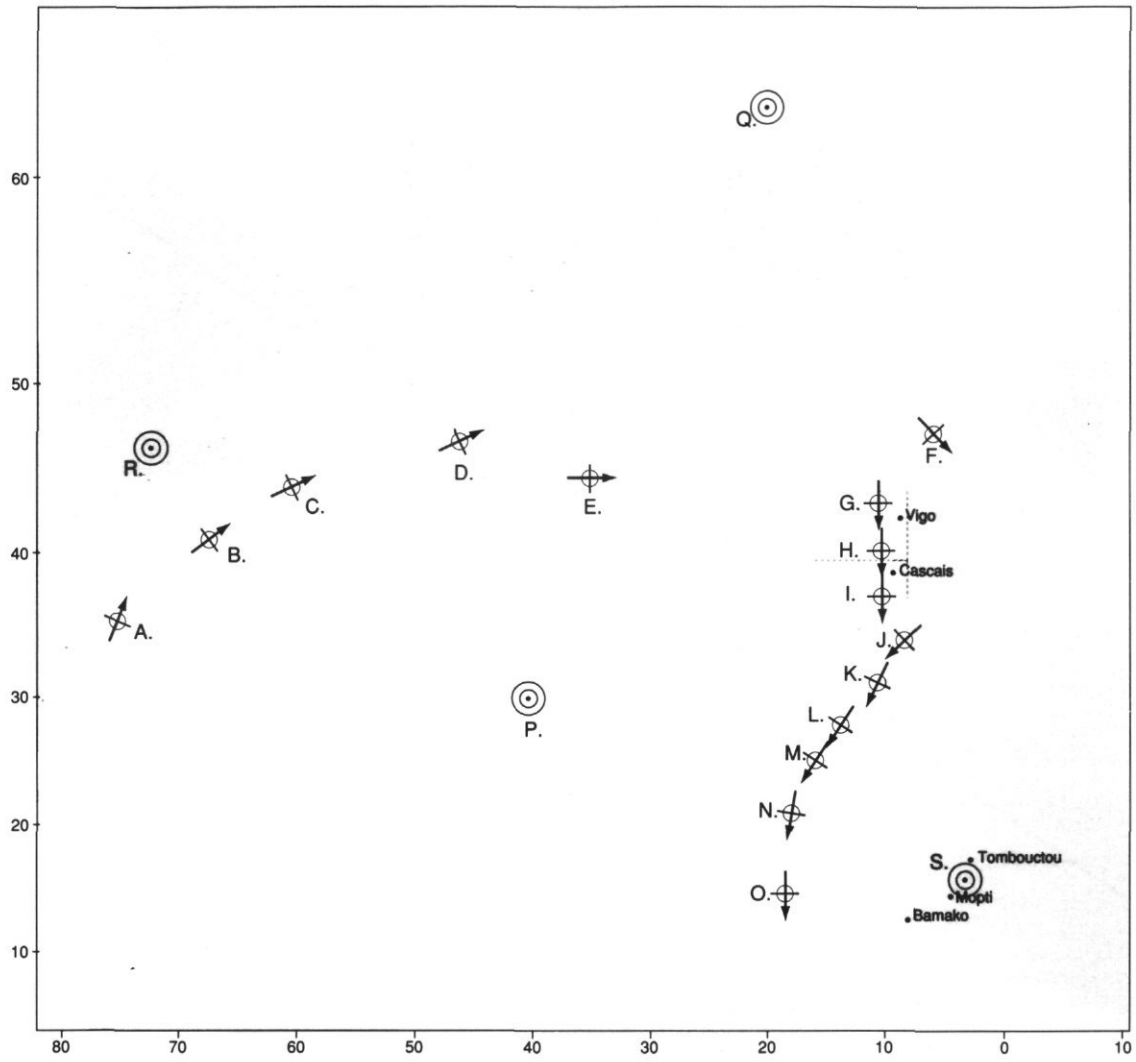

Figure 3. The North Atlantic. Locations at which time series have been constructed are indicated by the symbols (identified by capital letters to correspond to the time series plots shown in Figs. 4, 6, and 7 and to the entries in Tables 1 to 3). Arrow symbols (A to O) indicate positions of time series of stress estimates and point in the direction toward which the components treated in Figure 4 and Table 1 have been resolved. Circular symbols (P to S) indicate positions of atmospheric pressure time series treated in Figure 6 and Table 2.

performed (Table 1): (1) for the entire period, (2) for the period 1958–1973 of particularly rapid increases in several systems examined by Bakun (1990), and (3) for the period since 1973.

On the western side of the Atlantic the long-term (1946–1990) trend toward increasing alongshore windstress is highly significant at Cape Hatteras (A). The trend remains positive but becomes progressively weaker northward along the North American continental boundary (B, C), where the orientation of the coast is such that the perpendicular inland direction becomes increasingly northward, thus tending to oppose continental heating in the interior relative to the ocean area. Along the northern limb of the Gyre, where direct contact with a continental landmass is lost, the trend actually turns negative (D, E). Approaching the European continent, the negative trend diminishes to be nearly "flat" at the French coastal location (F). However, on the relatively meridional coast of the Iberian Peninsula (G, H, I) a strong, highly significant increasing trend is evident. The increase is particularly pronounced at the central location (H) where the "temperature contrast" effect of the Iberian landmass would tend to act most directly. Near the separation between the European and African continents, where the configuration of a "continental interior" is somewhat complicated by the presence of the Mediterranean Sea, etc., the increasing trend disappears (J). Note that a similar

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Erratum

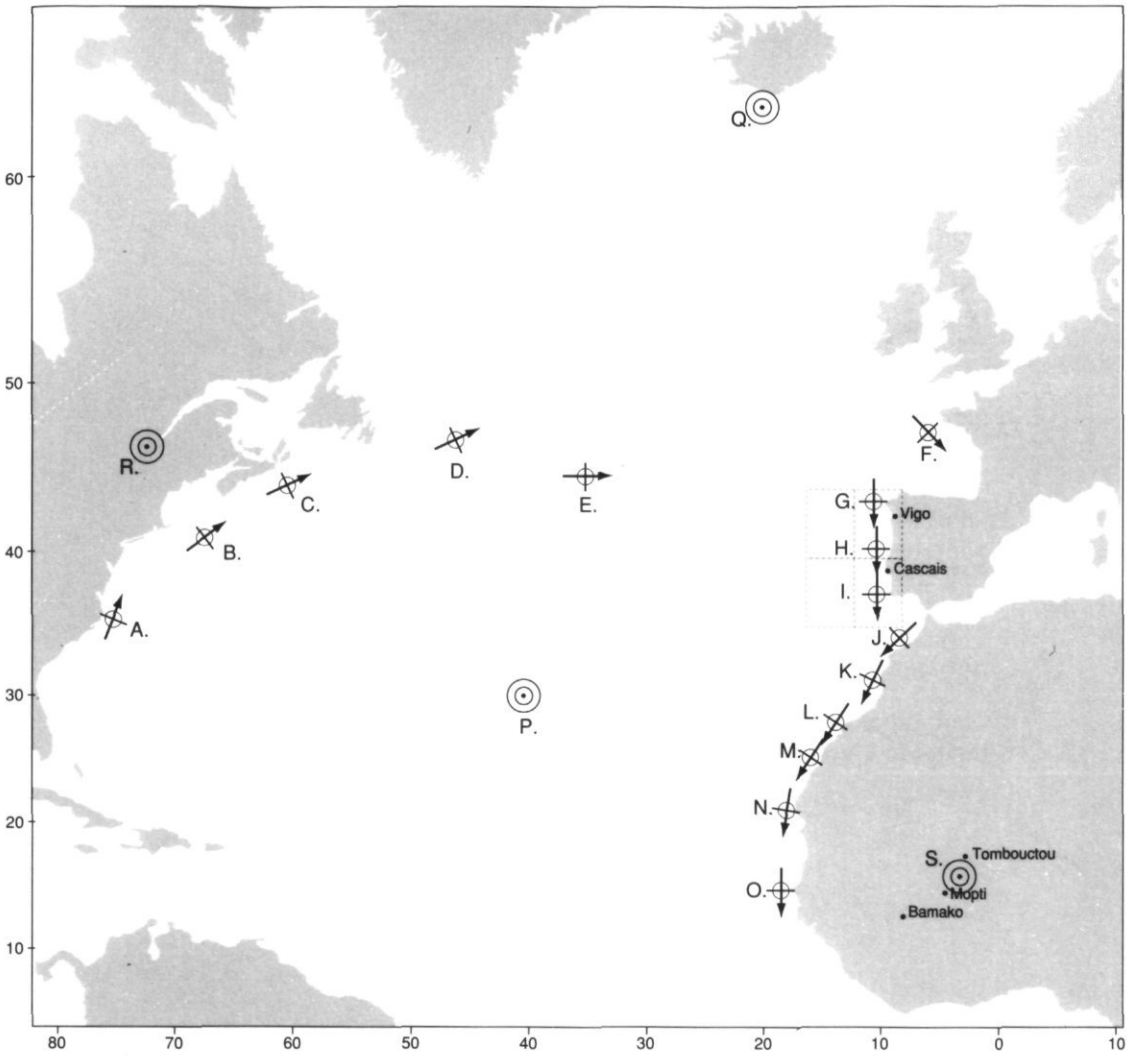
Hydrobiological Variability in the ICES Area, 1980–1989
ICES mar. Sci. Symp., 195: 316–325

Bakun, A. Global greenhouse effects, multi-decadal wind trends, and potential impacts on coastal pelagic fish populations

It is regretted that, because of a printing error, Figure 3 on page 319 of the above volume was reproduced without the coastline.

An amended page is printed on the reverse of this sheet for insertion at the appropriate point.

R. R. Dickson
Editor-in-Chief



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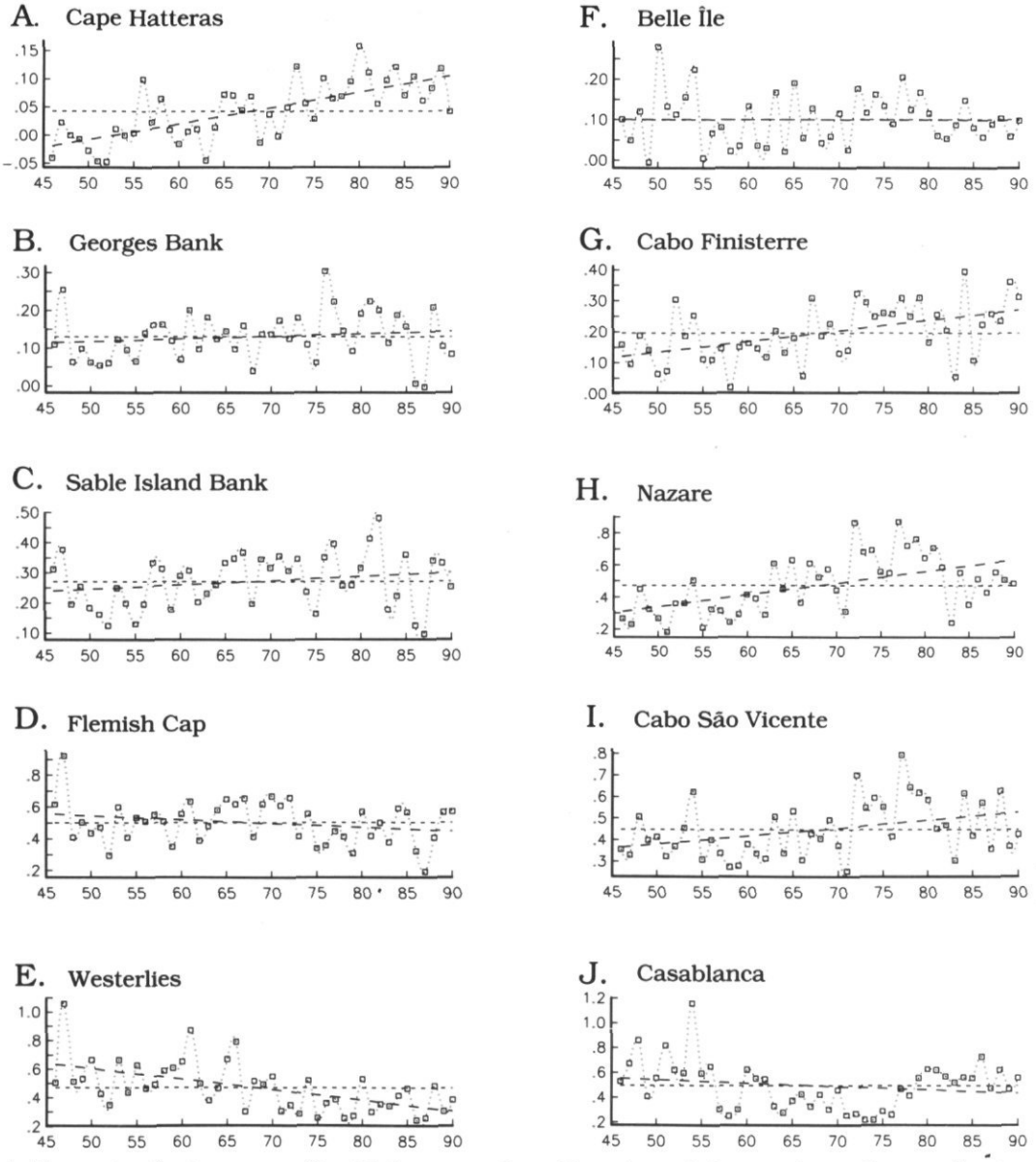


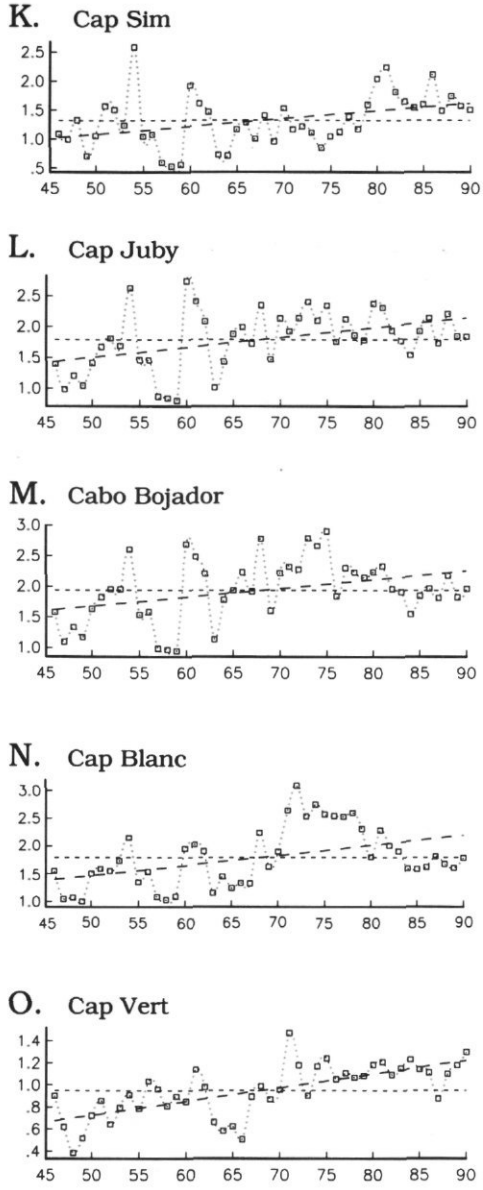
Figure 4. Time series of spring–summer (Oct–Mar) averages of monthly resultant windstress estimates (dyne cm^{-2}) at locations indicated in Figure 3. The long-term mean of each series is indicated by the horizontal line of short dashes. The linear trend, fitted by least squares, is indicated by the line of longer dashes.

disappearance of the increasing trend off western North America occurs off the Baja California Peninsula, where the Gulf of California occupies what would be the "continental interior" (Bakun, 1990). The long-term increasing trend reappears strongly to the south along the African coast (K through O), where first the Sahara Desert and then the African savanna lands occupy the "interior".

The global pattern of steepening of the trend in the period 1958–1973 followed by a leveling off or reversal in the later period 1973–present, apparent in the series presented by Bakun (1990) and also by Roy (1990), is evident in many of these series (see columns located to the right in Table 1).

Are the trends real?

The bulk of the available wind reports in the historical files are Beaufort estimates rather than anemometer measurements. These estimates, which are based on visual effects of winds on the sea surface, are "raised" to reflect a 10 m "mast height" reference using standard conversion tables. However, since the 1950s, the fraction of anemometer measurements in the marine report



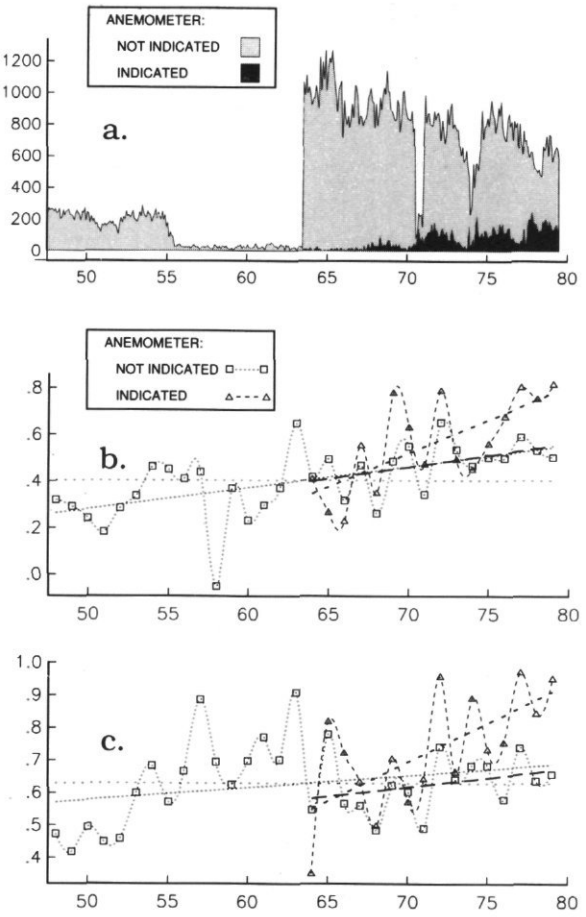


Figure 5. Time series constructed from samples of maritime reports (from within 5° areal quadrangles indicated in Figure 3 by fine dashed lines off the Iberian Peninsula) which have been "split" according to the "measured or estimated" indicator in the report format. (a) Numbers of available reports per month in the northeastern quadrangle (off Vigo). (b) April to September means of windstress estimates (dyne cm^{-2}) in the northeastern quadrangle. Shorter heavy dashes indicate the linear trend of the anemometer-based series. Longer heavy dashes and fine dotted line indicate the linear trend of the Beaufort estimate-based series over the same 1964–1979, and the longer 1948–1979, periods respectively. (c) April to September means of windstress estimates in the southeastern quadrangle (off Cascais).

files has been increasing (Ramage, 1987). It appears that the majority of ship's anemometers are actually installed at greater heights above the sea surface than the 10 m standard height to which the Beaufort estimates are calibrated (Cardone *et al.*, 1990). This would cause an overestimate of the wind at 10 m which would lead to a corresponding overestimate of the sea-surface stress. Thus there may be an artificial increasing trend in wind speeds inferred from the maritime data files.

For example, Whysall *et al.* (1987) report an apparent strengthening of zonal winds of about 1 m sec^{-1} from 1961 to 1983 over a large region of the tropical Pacific Ocean. However, Wright (1988) and Posmentier *et al.* (1989) conclude that such a trend is inconsistent with a lack of evidence of corresponding changes in sea-surface temperature and sea-level patterns and therefore must be spurious. Note that in upwelling regions, the "anemometer effect" would be amplified because the particularly stable lower atmosphere in these regions would cause larger differences between the wind directly at the sea surface, which is reflected in a Beaufort estimate, and the wind at mast height which would be measured by an anemometer. (Moreover, it is possible that even the historical series of analyzed fields of atmospheric pressure could be effected. In most analysis schemes, both in the recent objective computer-based methodologies and in the earlier subjective manual methodologies, some

Table 1. Trends in April–September averages of monthly resultant stress components oriented as indicated in Figure 2. (Bold numbers: slope of the linear trend line, dyne cm^{-2} yr^{-1}. In parentheses: correlation with a pure linear trend. Significance: * = $p < 0.05$; ** = $p < 0.01$. Indicated significance levels not adjusted for residual autocorrelation.)

		1946–1990	1958–1973	1973–1990
A.	Cape Hatteras	**0.0028** (0.72)**	**0.0018** (0.28)	**0.0001** (0.01)
B.	Georges Bank	**0.0007** (0.14)	**0.0014** (0.21)	**−0.0050** (0.33)
C.	Sable Island	**0.0014** (0.21)	**0.0066** (0.56)**	**−0.0030** (0.15)
D.	Flemish Cap	**−0.0024** (0.24)	**0.0048** (0.30)	**0.0028** (0.13)
E.	Westerlies	**−0.0074** (0.56)**	**−0.0106** (0.42)*	**0.0007** (0.04)
F.	Belle Isle	**−0.0010** (0.01)	**0.0001** (0.01)	**−0.0042** (0.53)*
G.	Cabo Finisterre	**0.0034** (0.50)**	**0.0049** (0.39)	**−0.0001** (0.00)
H.	Nazare	**0.0073** (0.54)**	**0.0167** (0.62)**	**−0.0161** (0.57)**
I.	Cabo San Vicente	**0.0037** (0.38)**	**0.0044** (0.23)	**−0.0098** (0.42)
J.	Casablanca	**−0.0027** (0.19)	**−0.0209** (0.60)**	**0.0197** (0.70)**
K.	Cap Sim	**0.0134** (0.39)**	**−0.0061** (0.08)	**0.0394** (0.56)**
L.	Cap Juby	**0.0160** (0.44)**	**0.0287** (0.31)	**−0.0188** (0.39)
M.	Cabo Bojador	**0.0144** (0.37)**	**0.0389** (0.40)	**−0.0480** (0.71)**
N.	Cap Blanc	**0.0182** (0.45)**	**0.0465** (0.51)*	**−0.0709** (0.89)**
O.	Cap Vert	**0.0124** (0.69)**	**0.0033** (0.24)	**0.0043** (0.22)

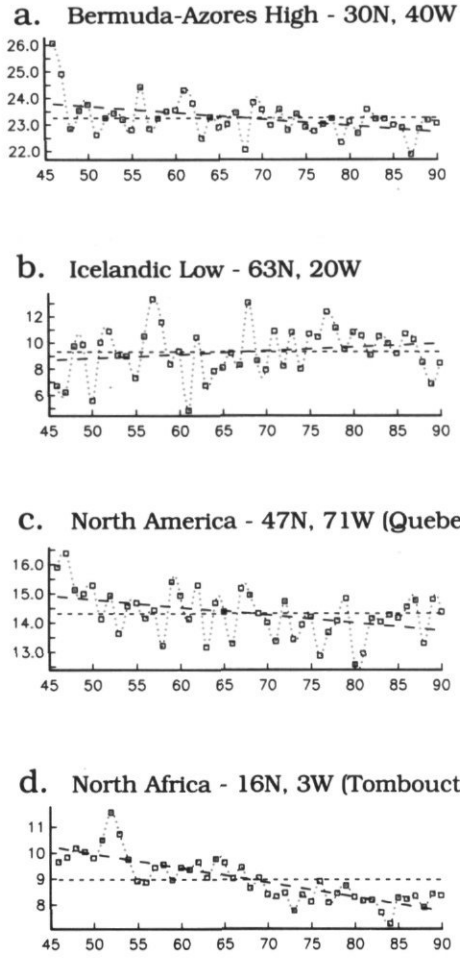


Figure 6. Spring–summer (Apr–Sep) time series of atmospheric pressure (mb-1000), produced at locations P to S indicated in Figure 3 from large-scale analyzed pressure fields (source: US Navy, Fleet Numerical Oceanography Center).

provision has been made to ensure that the synoptic pressure gradients reflect a reasonable dynamic balance with contemporaneous wind reports.)

Measured versus estimated time series

In the coded data record format in which the maritime reports are archived, there is one space which is coded "blank" if the wind report is a Beaufort estimate and "zero" if it is an anemometer measurement. However, there is no allowance for missing information, which apparently would be coded "blank". Thus it is likely that some anemometer measurements may be mistaken for Beaufort estimates in the files. There is also the possibility of erroneous codes in certain of the source decks (Cardone *et al.*, 1990).

When the available report file is divided according to the "measured or estimated" indicator, it is clear that the proportion of anemometer measurements has indeed increased in the North Atlantic area since the early 1960s (Fig. 5a). But when separate time series are produced from the split samples, both the measured and estimated windstress time series tend to separately corroborate the upward trend (examples are shown for two separate areas in Figures 5b and c). Note that after about 1968, when there began to be substantial numbers of reports in the measured series, the year to year fluctuations, as well as the upward trend, tend to be "mirrored" in both the measured and estimated time series. However, the estimates in the measured series tend to be consistently higher than in the estimated series. Thus the indication is that there is an "anemometer effect" that may indeed account for a minor increasing trend in wind series produced from maritime wind reports, but that this accounts for only a fraction of the increasing trend off the Iberian Peninsula (and assumedly in other regions of equal or stronger indicated trend).

Table 2. Trends in April–September average surface atmospheric pressures, extracted from large-scale analyzed fields, at locations indicated in Figure 2. (Bold numbers: slope of the linear trend line, mb yr^{-1}. In parentheses: correlation with a pure linear trend. Significance: * = $p < 0.05$; ** = $p < 0.01$. Indicated significance levels not adjusted for residual autocorrelation.)

		1946–1990	1958–1973	1973–1990
P.	Bermuda High	**−0.0242** (0.45)**	**−0.0161** (0.18)	**−0.0074** (0.10)
Q.	Icelandic Low	**0.0286** (0.20)	**0.0044** (0.01)	**−0.1195** (0.48)*
R.	N. America (Quebec)	**−0.0264** (0.43)**	**−0.0132** (0.12)	**0.0511** (0.40)
S.	N. Africa (Tombouktou)	**−0.0554** (0.82)**	**−0.0714** (0.69)**	**−0.0079** (0.11)

Corroboration in barometric pressure trends

Atmospheric pressure time series at four locations (marked P, Q, R, and S in Fig. 3) have been examined for "keys" to the increasing spring–summer wind trends identified in the North Atlantic. But rather than increasing in intensity in recent decades, both the major oceanic high and low pressure centers, the "Bermuda–Azores High" (P) and the "Icelandic Low" (Q) areas, seem to have been relaxing (Fig. 6a,b; Table 2). However, the seasonal continental low pressure systems (R and S) have been increasing in intensity (Fig. 6c,d; Table 2), in accord with the greenhouse scenario. The thermal low in the interior of Northwest Africa (S) appears to have been intensifying in a particularly dramatic manner, in congruence with the strongly increasing wind trends along the West African coast (Fig. 4, K–O, Table 1).

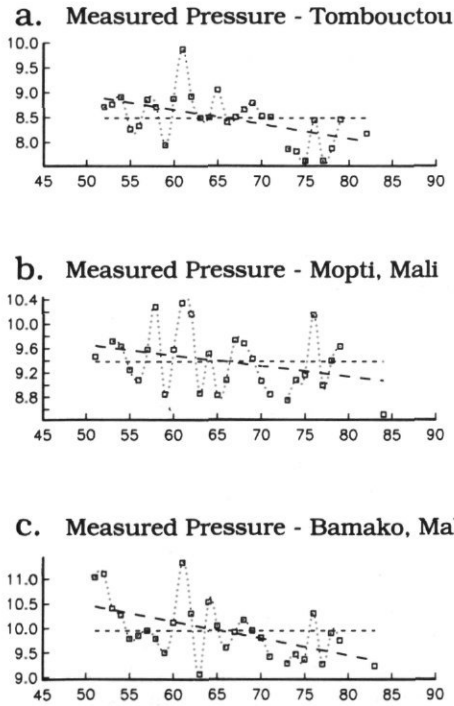


Figure 7. Spring–summer (Apr–Sep) time series of independently measured barometric pressure (mb-1000) at three stations in Northwest Africa (Fig. 3).

The trends shown in Figure 6 are produced from large-scale analyzed pressure fields. Data in the Northwest African interior are rather sparse and changes in data pattern and analysis procedures might be suspected as being involved in the indicated trend. However, several actual measured series (Fig. 7) from a group of stations in North Africa appear in each case to independently substantiate the trend.

On a more local scale, the geostrophic balance of the apparent increase of the equatorward windstress in the data-rich area off the Iberian Peninsula was investigated, not with analyzed fields of pressure data which might conceivably have been contaminated by an "anemometer effect", but with simple averages of maritime barometric pressure reports. Available reports within each of the four 5° areal quadrangles outlined by fine dashed lines in Figure 3 were averaged by month. The differences in the monthly mean values in each zonally arranged pair of quadrangles (i.e. the northern and southern pairs respectively) were computed as an index of the monthly mean zonal pressure gradient. Note that by sampling sporadically distributed (in time and space) pressure observations at each "end" of a desired gradient rather than paired observations taken simultaneously and in consistent locations, one samples a population of apparent gradients far more extensive than could actually physically exist. This results in a very "noisy" estimate of a monthly mean gradient. Even so, the computed differences indicate a linear increase of mean equatorward geostrophic wind for the April–September period of 0.021 m s^{-1} yr^{-1} and 0.034 m s^{-1} yr^{-1} for the northern and southern pairs of quadrangles respectively.

An independent verification from within the ocean itself

Low frequency increases in upwelling-producing windstress should be reflected in corresponding increases in alongshore slope in mean sea level (Csanady, 1978; Enfield and Allen, 1980; Bigg and Gill, 1986). Bigg (1988) reported an increase in sea-level differences between sea-level gauge stations along the South American Coast that tends to corroborate the increase in Pacific tradewinds reported by Whysall *et al.* (1987). In the North Atlantic region, the Iberian Peninsula pro-

trudes out from the European continent and presents a coherent meridional stretch of coastline which may be somewhat isolated from interference from propagating coastal-trapped waves and thus an excellent site to look for a local sea-level response to the indicated spring–summer increasing trend in alongshore windstress at the coast (G, H, and I in Fig. 4 and in Table 1). Since we are looking for a spring–summer greenhouse effect, examining the difference between winter and summer slopes allows a filtering of long-term inhomogeneities due to repair, replacement or relocation of gauges, etc. Indeed, there has been a significant long-term increasing trend in the summer–winter difference in sea-level slope (1.78 mm yr^{-1}, $r = 0.56$, $p < 0.01$), corrected for atmospheric pressure effect, between Vigo, Spain, and Cascais, Portugal (Fig. 8). Lest one worry that the atmospheric pressure corrections which are based on analyzed fields may themselves have introduced the trend in corrected sea level, the increase in corresponding slope in uncorrected sea level is nearly as pronounced (1.64 mm yr^{-1}, $r = 0.48$, $p < 0.01$).

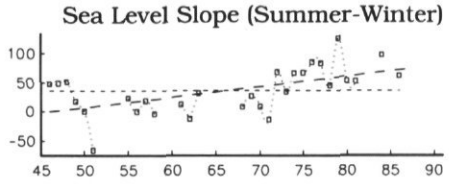


Figure 8. Seasonal (spring–summer minus fall–winter) difference in sea level slope along the west coast of the Iberian Peninsula (value at Cascais minus value at Vigo, measured in mm).

Fall–winter trends

For reference, corresponding cold season (Oct–Mar) averages of monthly resultant windstress estimates have also been produced for the same locations (Fig. 3) and examined for trends. Because the resultant wind directions during fall–winter vary more energetically than those during spring–summer, the trend analysis for fall–winter (Table 3) was performed simply in terms of scalar magnitude of the monthly resultant wind. Two regions of significant long-term increase in resultant windstress appear. One is for the mid-ocean location "E" (Westerlies). The other major area of fall–winter increase was along the West African coast (locations "K" to "O"). Here the increase was almost entirely in the alongshore component, as in the spring–summer situation (e.g. at the southernmost location "O", the trend in the alongshore component of the resultant monthly stress is actually larger and more significant than in its scalar magnitude, e.g. 0.0116 dyne cm^{-2} yr^{-1}, $r = 0.53$, $p < 0.01$). In the interior of tropical Africa there is a thermally enhanced continental low year round. Here the thermal effect, related to greenhouse warming, would be expected to act year round.

Thus if one looks only at the locations where the proposed greenhouse mechanism would not be expected to be effective in the fall–winter half of the year (i.e. excluding locations K to O where the greenhouse mechanism would tend to act year round) one actually finds, in the fall–winter period, more long-term decreases than increases in monthly resultant stress. This suggests that the artificial "anemometer problem" has not acted as the controlling factor in the multi-decadal spring–summer wind trends observed in the North Atlantic.

Table 3. Trends in October–March averages of scalar magnitudes (i.e. regardless of direction) of monthly resultant stress values. (Bold numbers: slope of the linear trend line, dyne cm^{-2} yr^{-1}. In parentheses: correlation with a pure linear trend. Significance: * = $p < 0.05$; ** = $p < 0.01$. Indicated significance levels not adjusted for residual autocorrelation.)

		1946–1991	1958–1973	1973–1991
A.	Cape Hatteras	**−0.0030** (0.42)**	**−0.0051** (0.30)	**0.0009** (0.08)
B.	Georges Bank	**−0.0001** (0.011)	**−0.0077** (0.42)*	**−0.0018** (0.07)
C.	Sable Island	**0.0009** (0.11)	**0.0006** (0.04)	**−0.0062** (0.27)
D.	Flemish Cap	**0.0031** (0.15)	**0.0024** (0.05)	**0.0035** (0.07)
E.	Westerlies	**0.0069** (0.30)*	**0.0054** (0.11)	**0.0122** (0.24)
F.	Belle Isle	**0.0023** (0.12)	**−0.0017** (0.04)	**−0.0015** (0.03)
G.	Cabo Finisterre	**−0.0035** (0.17)	**0.0051** (0.10)	**−0.0049** (0.12)
H.	Nazare	**−0.0033** (0.23)	**−0.0043** (0.13)	**−0.0030** (0.10)
I.	Cabo San Vicente	**−0.0005** (0.05)	**−0.0009** (0.04)	**−0.0018** (0.10)
J.	Casablanca	**−0.0005** (0.09)	**−0.0001** (0.01)	**0.0032** (0.26)
K.	Cap Sim	**0.0032** (0.45)**	**−0.0006** (0.04)	**0.0056** (0.36)
L.	Cap Juby	**0.0078** (0.56)**	**0.0073** (0.29)	**−0.0062** (0.25)
M.	Cabo Bojador	**0.0084** (0.48)**	**0.0179** (0.53)**	**−0.0232** (0.65)**
N.	Cap Blanc	**0.0115** (0.38)**	**0.0417** (0.59)**	**−0.0548** (0.79)**
O.	Cap Vert	**0.0078** (0.30)*	**0.0295** (0.44)*	**0.0077** (0.16)

Concluding remark: basin-scale relaxation versus intensification at the continental–ocean boundary

Simulations of "enhanced CO_2" scenarios using general circulation models (GCMs) have indicated that greenhouse warming of the lower atmosphere should be intensified toward higher latitudes (Manabe and Wetherald, 1980). In such a case, the global-scale latitudinal temperature gradient, as well as the intensity of the circulation tending to dissipate this gradient, should decrease. However, seasonal temperature contrasts between coastal ocean regions and the adjacent continental landmasses would be intensified. This would tend to feed back into strengthened coastal wind systems (Bakun, 1990). In this paper it has been noted that, during the spring–summer warming season at least, the major oceanic pressure systems of the Atlantic (the Bermuda–Azores High and the Icelandic Low) have indeed become less intense over the past several decades. Likewise, it has been noted that the continental "thermal" low pressure systems appear to have intensified, as have the associated coastal ocean wind systems. It will be this interaction of seasonally intensified "land boundary" effects with the expected relaxation of global-scale circulation that should determine, in each local instance, the particular wind-related effects of climate change on marine ecosystem processes and populations.

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Sensitivity of fish and other populations to environmental fluctuation

B. J. Rothschild

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The sensitivity of fish and other populations to environmental fluctuations is considered in terms of well-known dynamic population models. The slightest change in parameter values within certain ranges can evoke large qualitative and quantitative changes in population trajectories. Populations reduced by fishing, or anthropogenic substances, which compensate for reductions by changes in vital rates can easily transit among stable, periodic, and chaotic population dynamics.

B. J. Rothschild: Chesapeake Biological Laboratory, Solomons, Maryland 20688, USA.

Introduction

Fish-stock variability is usually studied in terms of its dependence upon stock size. This dependence is usually thought of in terms of the recruitment–stock relationship. Deviations from the relationship are usually considered as environmentally related statistical noise. However, characteristics of various non-linear dynamic models suggest that departures from traditionally used static models can arise from non-linear considerations, and for this reason we might not be surprised to find that a "poor" recruitment–stock relationship viewed in a static framework would actually be a "good" relationship taking account of the underlying dynamics of the model.

Specifying the sources of environmentally related statistical noise is one of the more important tasks in fishery research. Any identification or attribution of noise to specific environmental variables reflects an increase in understanding, and an increase in understanding reflects the capability to make better fishing-management decisions (see Rothschild and Heimbuch, 1983).

One of the difficult problems in assessing environmentally related statistical noise is that complex non-linear phenomena, the quintessential components of the population-dynamics process (see Rothschild, 1986, Chapter 8), make it difficult to separate population dynamics, environmentally induced fluctuations, and noise.

A simple-minded view of the exquisite complexity of the non-linear interactions in the population-dynamics process can lead one to believe that anthropogenic or naturally caused declines in fish populations result in compensation and that because of compensation populations increase, and hence the declines have little effect on the population. Another related view is that, because fish populations have a seeming superabundance of eggs, any artificial or environmentally caused reductions in the numbers of eggs will have little effect on the population.

Some of these assertions are obviously not true, as can be seen from simple theory. For example, recruitment–stock theory – at least as it is used in stock assessment (i.e. plotting recruitment against stock) – readily shows that any change in stock will be met with a measurable *average* change in recruitment. Likewise, any change in stock size that drives changes in natural mortality rates and growth will result in a change in biomass that may result in either an increase or decrease in recruitment.

These readily computed changes are related more to population kinematics than to population dynamics. A study of dynamics reflects that the changes in populations are complex. While the nature of the changes is predictable under certain circumstances, a fuller understanding of biological mechanisms is required before the dynamics effects can be separated from the environment and from statistical noise.

This paper reviews two dynamics problems and one problem in population statistics. The first dynamics problem is the prototypical recruitment–stock relationship treated as a one-dimensional logistic map. The second dynamics problem involves multiple species predator–prey interactions; the third problem reviews sensitivity to early life-history mortality.

The one-dimensional logistic map

The one-dimensional logistic map is an ideal device for drawing our attention from kinematic stock-assessment recruitment–stock relationships to population-dynamic recruitment–stock relationships. The stock-assessment approach has generated a number of curve-fitting exercises that consider only the relations between recruitment and stock and not the dynamics of recruitment–stock–recruitment–stock–recruitment, etc. (Rothschild and Mullen (1985) have presented a stochastic, nonparametric *dynamic* recruitment–stock model).

In addition to being important to the dynamics of populations, the subject is of considerable historic importance inasmuch as both Ricker (1975) and Beverton and Holt (1957) in particular, clearly recognized "chaos" (Fig. 1) long before it had a name and became a subject of great popularity.

All recruitment–stock theory stems from the notion that recruitment results from a parent stock. This idea can be formulated in a number of ways (of which the Ricker model and the Beverton and Holt model are but special cases).

Generally, however, we can write,

$$N_{t+1} = N_t F(N_t) = f(N_t) \tag{1}$$

with the steady state, $N_t = N_{t+1} = N^*$.

So from Equation (1), we have

$$N^* = N^* F(N^*) = f(N^*). \tag{2}$$

Hence the steady state is attained when $N^* = 0$ or $F(N^*) = 1$. As is well known (see Murray, 1989), the dynamic behavior of the system depends upon its eigenvalue at the equilibrium point,

$$\lambda = f'(N^*) = \left[\frac{df(N_t)}{dN_t}\right]_{N_t = N^*} \tag{3}$$

Near the steady state N^*, we have the following conditions; if,

$-1 < \lambda < 0 \Rightarrow N^*$ is stable with decreasing oscillation
$\lambda = -1 \Rightarrow N^*$ is neutrally stable
$\lambda < -1 \Rightarrow N^*$ is unstable with growing oscillation

Now let us examine a specific form of the recruitment–stock relationship, the logistic model.

$$N_{t+1} = rN_t\left(1 - \frac{N_t}{K}\right) \tag{4}$$

where N_{t+1} can be thought of as recruitment and N_t as the stock, r is the "rate of increase", and K is the "carrying capacity".

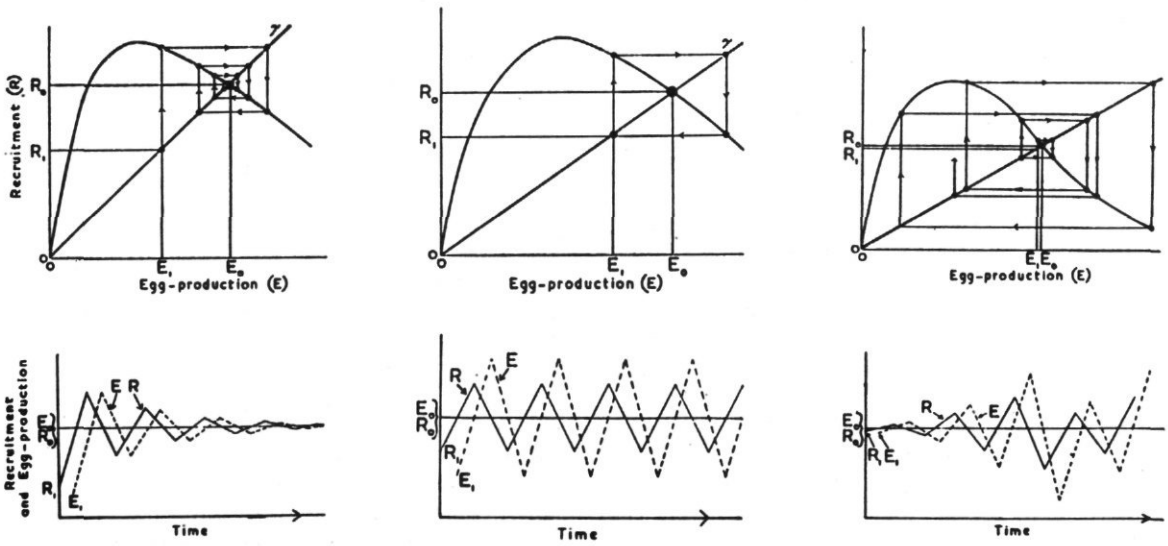

Figure 1. In 1957 Beverton and Holt demonstrated the onset of chaos in the recruitment–stock relationship: "Figs. 6.6 the Dynamics of a Self-Regenerating Model with the Egg–Recruit Relationship . . . the upper part of each diagram shows successive values of recruitment and egg production following a disturbance from an initial steady state, while the lower part shows these changes plotted as time series. Fig. 6.6.1 Slope of recruitment curve shallower than that of egg-production curve at intersection. Recovery of initial steady state following a transitory disturbance. Fig. 6.6.2 Slope of recruitment curve equal to that of egg-production curve at intersection. Permanent oscillations of equal amplitude following a transitory disturbance. Fig. 6.6.3 Slope of recruitment curve steeper than that of egg-production curve at intersection. Permanent oscillations of increasing amplitude following a transitory disturbance." (From Beverton and Holt, 1957).

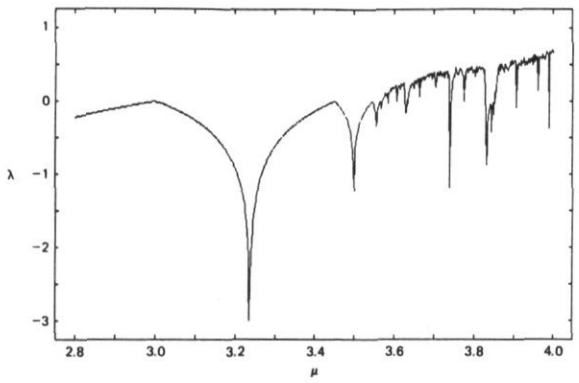

Figure 2. Lyapunov exponent λ versus μ for the logistic map. Sensitivity to initial conditions occurs where the exponent is positive (based on Baker and Golub, 1990).

Setting K = 1 rescales Equation (4) to

$$u_{t+1} = ru_t(1 - u_t). \tag{5}$$

Equation (5) is then both a recruitment–stock relationship and the logistic map.

As implied by Beverton and Holt, the nature of the map varies to a considerable degree depending upon the magnitude of λ or r. This is easily discerned from the map's bifurcation diagram, which shows stability, chaos, and periodic windows. Further insights can be obtained from considering the map's Lyapunov exponent (Fig. 2) and entropy content (Fig. 3) as a function of its intrinsic rate of increase.

The main point to be derived from the analysis is that minute changes in r can generate major changes in the dynamic behavior of the population – changes from stability, to periodicity, to chaos.

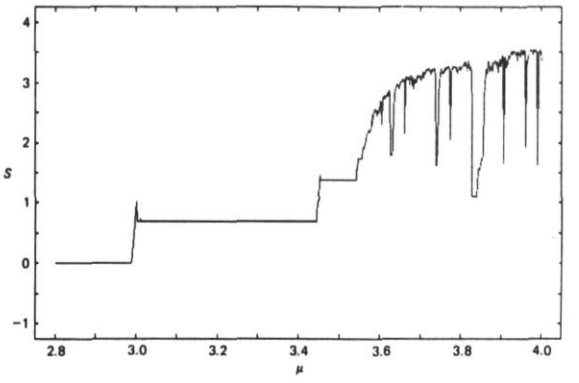

Figure 3. Entropy S as a function of μ for the logistic map. The maximum entropy corresponding to equal probability for each of the 40 cells is 3.6888 (based on Baker and Golub, 1990).

Multiple species interaction and neighborhood stability analysis

Let us consider a system which in some ways is less mathematically complex, but perhaps biologically more complex. Consider two populations, a predator and a prey, for example, where each population might be represented by equations similar to Equation (4), but with the addition of population interactive terms. Following the notation and analysis of Nisbett and Gurney (1982), these populations can be represented by the differential equations:

$$\begin{aligned} \dot{N}_1 &= G_1(N_1,N_2) \\ \dot{N}_2 &= G_2(N_1,N_2) \end{aligned} \tag{6}$$

Now we make a gross simplification by transiting from the non-linear characteristics of Equation (6) to the much simpler local-linear analysis. First we find the steady state,

$$\begin{aligned} G_1(N_1^*,N_2^*) &= 0 \\ G_2(N_1^*,N_2^*) &= 0, \end{aligned} \tag{7}$$

define perturbations about the steady states,

$$\begin{aligned} n_1 &= N_1 - N_1^* \\ n_2 &= N_2 - N_2^*, \end{aligned} \tag{8}$$

and expand about the steady state,

$$\begin{aligned} G_1(N_1,N_2) &\simeq G_1(N_1^*,N_2^*) + \left(\frac{\partial G_1}{\partial N_1}\right)^* n_1 + \left(\frac{\partial G_1}{\partial N_2}\right)^* n_2 \\ G_2(N_1,N_2) &\simeq G_2(N_1^*,N_2^*) + \left(\frac{\partial G_2}{\partial N_1}\right)^* n_1 + \left(\frac{\partial G_2}{\partial N_2}\right)^* n_2, \end{aligned} \tag{9}$$

to obtain first-order constant coefficient differential equations

$$\begin{aligned} \dot{n}_1 &= A_{11}n_1 + A_{12}n_2 \\ \dot{n}_2 &= A_{21}n_1 + A_{22}n_2. \end{aligned} \tag{10}$$

We can see that the coefficients of Equation (10) comprise the 2 × 2 matrix **A**. The properties of **A** specify the dynamic behavior of the trajectories of **n**. These are classified in Figure 4. Figure 4 shows that there are at least seven regions for which the determinant and trace of **A** elicit different dynamic behaviors – the stable and unstable mode; the stable and unstable spiral; periodic; and complicated saddle-point behavior. Thus, in common with the logistic map, even the simplified Equation (10) possesses complex behavior and in some circumstances even the slightest change in parameters can generate large changes in population variability.

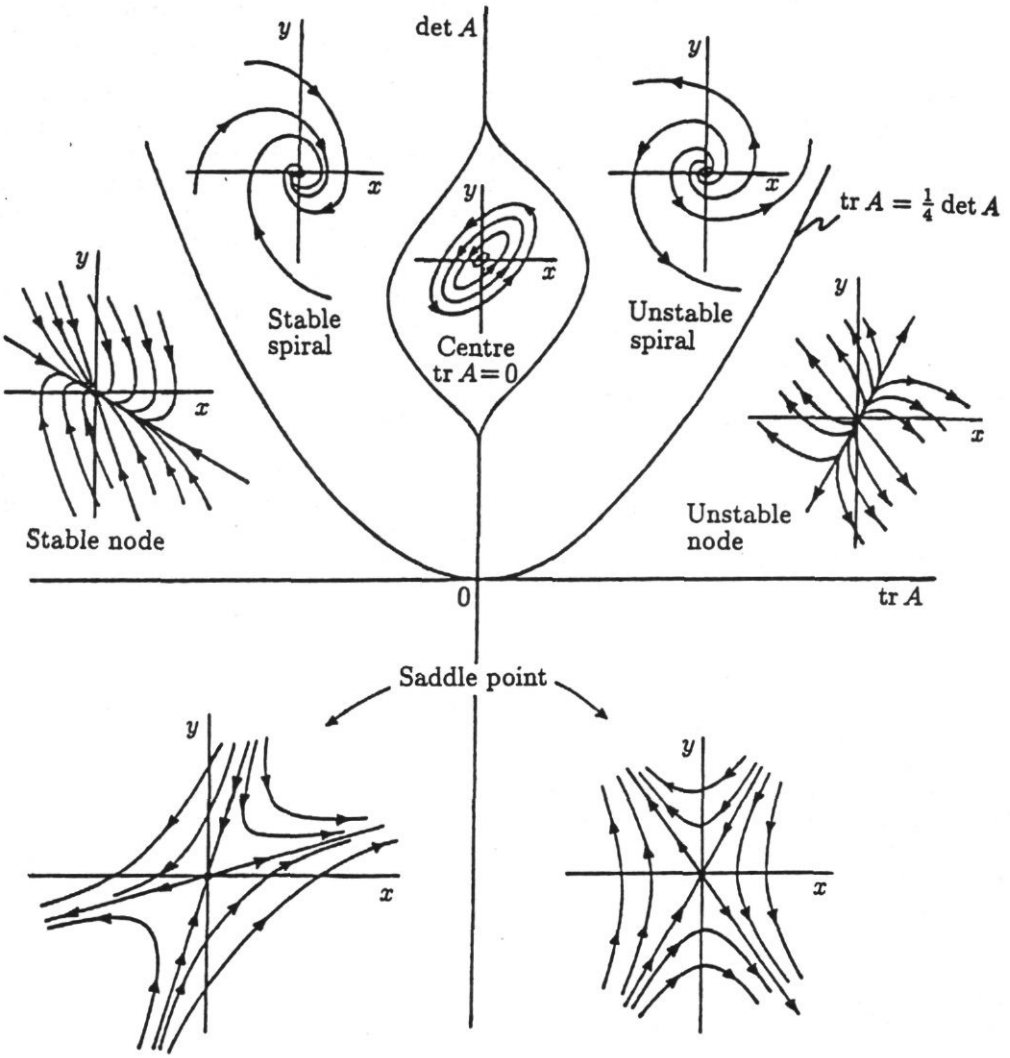

Figure 4. Regions of differing dynamical behavior (based on Murray, 1989).

Sensitivity of traditional mortality

Beyer (1989) has brought together many of the basic ideas on how variability in life history parameters might affect recruitment. As a point of departure he noted that a female cod spawned in the order of 10^6 eggs. For a cod population to maintain a steady state it would then have to have a survival rate in the order of 10^{-6}. Of course, different species of fish have different fecundity and hence different survival rates to the time or index age equivalent to the reproductive potential of each female. For example, the life-time reproductive potential fecundity of a Pacific herring is about 6.5×10^4 and of a pink salmon about 3×10^3. The results of arranging these fecundities in terms of a pair of spawners and computing instantaneous rates, corresponding to the survival rates, are illustrated in Table 1.

The values in Table 1 are of course very rough guesses, but they do show that the mortality rates in the early life history are considerably in excess of the mortality rates in the adult stages. Beyer considered the consequences of small perturbations in the mortality rates, 10%, for example. So for some fixed interval of index age, we would have a decreased number of individuals in the population:

$$N_D = e^{-1.1m} \tag{11}$$

and an increased number of individuals in the population

$$N_I = e^{-0.9m} \tag{12}$$

Table 1.

Species	Female per fecundity	Per pair survival rate S	M	Index age	Average instantaneous rate per year	$S^{-0.2}$
Cod	10^6	2×10^{-6}	13	4	3.25	14
Pacific herring	6.5×10^4	3×10^{-3}	6	7	0.9	3
Pink salmon	3×10^3	7×10^{-2}	3	2	1.5	2

So the ratio of the largest population size to the smallest population size is

$$\frac{N_I}{N_D} = \frac{e^{-0.9m}}{e^{-1.1m}} = e^{0.2m} = S^{-0.2} \qquad (13)$$

Application of the factor of $S^{-0.2}$ to the per pair survival rates in Table 1 then gives the ratio of the largest to smallest population size given a ±10% change in mortality. This shows that fish with the highest fecundity are subject to much greater swings in recruitment owing to the same small changes in mortality rate. This has many implications. Two are particularly important. The first is the capability of techniques for detecting changes in mortality rates of less than ±10%. The second is that because populations vary at about roughly the predicted range of variation, events associated with stabilization are probably more important than the specific causes of mortality.

Discussion

We have demonstrated that the parameters in standard models reveal a considerable sensitivity, particularly in certain parameter ranges. This sensitivity is of two types. In the non-linear models we were considering sensitivity in terms of the reproductive process. In the examination of mortality rates we were basically considering intra-year class effects.

We have shown that not only are stocks of fish sensitive to environmental factors that affect parameter values but that arguments that high-fecundity fish and "compensation" negate natural or anthropogenic factors are not well founded.

On the one hand, we can see that recruitment variability can be much larger in high-fecundity fish than in low-fecundity fish, given the same percentage change in mortality, and even more strikingly well-known compensatory changes which must affect r, the rate of increase, by increasing r at low population levels can lead to transitions among stable, oscillatory, and chaotic regimes where the transition can be triggered by the most minute change.

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ICES mar. Sci. Symp., 195: 331–348. 1992

Regional variability in fish–prawn communities and catches in the Barents Sea, and their relationship to the environment

E. M. Nilssen and C. C. E. Hopkins

Nilssen, E. M., and Hopkins, C. C. E. 1992. Regional variability in fish–prawn communities and catches in the Barents Sea, and their relationship to the environment. – ICES mar. Sci. Symp., 195: 331–348.

The Barents Sea comprises a northern area dominated by Arctic watermasses and a southern area dominated by Atlantic watermasses separated by a transitional Polar Front. The period 1980–1989 was characterized by large variability in the advection of Atlantic water, with strong influxes and elevated temperatures in 1982/1983 and again in 1989, and colder temperatures in the intermediate years. Strong and moderate recruitment of cod and haddock were associated with the elevated temperatures and the extension of the Atlantic domain northeastwards. In the cold periods poor recruitment of cod and haddock occurred, with relative and absolute increases in numbers of polar cod. Cluster analysis showed that the connectance and integrity of the fish–prawn communities varied with temperature, although the cod–haddock community of the Atlantic-dominated domain and the polar-cod-dominated community of the Arctic domain were relatively robust. Capelin, however, were essentially "boreo-Arctic", as they were transitional between the two. Prawns (*Pandalus borealis*) showed less variability than either cod–haddock or polar cod. Canonical correspondence analysis (CCA) involving species and the influence of "environmental" factors (e.g. latitude and longitude, bottom depth, survey year, and temperature) on their abundances and distributions in eight different areas of the Barents Sea showed that it was the northeastern areas which underwent most variability. These areas are clearly affected by variability in the position of the Polar Front, and thus dominance by either warmer or colder waters. CCA showed that temperature was the major environmental forcing function affecting the species gradient, although latitude and longitude were also important as they were covariates with temperature. The results are discussed with regard to historical data about stock variability in the Barents Sea, and suggest that herring prior to their collapse in the early 1970s were part of the cod–haddock community. The capelin, owing to their transitional position between Atlantic and Arctic communities, thus probably did not fully replace herring as suitable prey for cod, as this predator–prey cycle will be temporally and spatially displaced.

E. M. Nilssen and C. C. E. Hopkins: Department of Aquatic Biology, Norwegian College of Fisheries Science, University of Tromsø, N-9037 Tromsø, Norway.

Introduction

The Barents Sea is one of the highly productive continental shelf ecosystems (Fig. 1) and an important nursery and feeding area for several major commercial fish and shellfish stocks (see Postma and Zijlstra, 1988; Dragesund and Gjøsæter, 1988). Prior to the collapse of the Atlanto-Scandian herring (*Clupea harengus*) about 1970, the Barents Sea has traditionally acted as a nursery area for young herring (Dragesund and Gjøsæter, 1988). The Barents Sea has long supported considerable fisheries (see Hjort, 1914), and total landings of fish have ranged from about 1.5 to 4.5 million tonnes annually between 1950 and 1980 (Dragesund and Gjøsæter, 1988; Anon., 1990). The demise of the Atlanto-Scandian herring saw the establishment of a lucrative capelin (*Mallotus villosus*) fishery in the Barents Sea, with nominal total catches between 1977 and 1983 varying between 2.9 and 1.5 million tonnes, before declining to 123 000 tonnes in 1986 followed by a moratorium until the end of the decade (Hamre, 1985; Anon., 1990). The total catch of the Northeast Arctic (Barents Sea) stock of cod (*Gadus morhua*), close to a million tonnes in the mid-1970s, varied between 300 000 and 500 000 tonnes in the 1980s (Dragesund and Gjøsæter, 1988; Anon., 1990). Both the fisheries resources and the landings sustained from the Barents Sea have fluctuated widely in the past as well as in more recent times (Helland, 1908;

Helland-Hansen and Nansen, 1909; Hjort, 1914; Anon., 1948; Rollefsen, 1954; Ervik and Roman, 1975; Bergstad *et al.*, 1987; Loeng, 1989a).

The Barents Sea (Fig. 1) is a neritic, shelf-sea with a mean depth of about 230 m, covering an area of about 1.5 million km^2 (Zenkevitch, 1963). It is bordered to the north by the archipelagos of Spitsbergen and Franz Josef Land, to the east by Novaya Zemlya, and to the south by the mainland of Norway and the USSR. Westwards the Barents Sea is delineated by the shelf break bordering the Norwegian Sea. The northern part of the sea is covered by annually formed sea-ice in the winter, and is primarily under the influence of cold watermasses of Arctic origin flowing from the north. Atlantic water flows into the Barents Sea from the southwest, occupying the area south of the Polar Front, the latter being a region in which the two watermasses meet (Tantsiura, 1959; Loeng, 1989b). In the west the Front is conspicuous, delineating two distinct regimes, while in the east it is diffuse as extensive mixing between the watermasses occurs (Loeng, 1989b). The variability of Atlantic inflow into the Barents Sea may be great (Worthington, 1970; McClimans and Nilsen, 1990), resulting in marked changes in sea temperature and ice cover, and the extent of the Atlantic and Arctic "domains" (Loeng, 1989a). This is considered also to have consequences for recruitment and distribution of several fish species in the Barents Sea (Ponomarenko, 1973; Sætersdal and Loeng, 1987; Nakken and Raknes, 1987; Shevelev *et al.*, 1987; Hamre, 1988; Loeng, 1989a; Cushing, 1990a).

It has become increasingly clear that periodic changes in the marine climate, frequently connected with baroclinic variation and surges in advected watermasses, significantly affect recruitment of fish stocks and the production cycles in general (Garrod and Colebrook, 1978; Cushing, 1982, 1990a, b; Hollowed *et al.*, 1987; Shepherd *et al.*, 1984; Dickson *et al.*, 1988; Laevastu *et al.*, 1988; Southward *et al.*, 1988). In addition to providing temperature "anomalies" at a given locality, changes in optimum environmental characteristics for specific species/communities may result in increases or decreases in suitable habitat areas or domains (Shuter and Post, 1990). Although a good deal of information exists regarding basic trends in climatic variability, including aspects of periodicity, and its influence on recruitment and distribution of several commercially important fish species in the Barents Sea (see previous paragraph), there is an almost total lack of information concerning community perspectives.

The present paper establishes and elucidates the cohesiveness of the commercially important demersal communities of deep-water prawn (*Pandalus borealis*) and fish in eight large "areas" of the Barents Sea between 1980 and 1989, a period characterized by clear fluctuations in the inflow of Atlantic water (Loeng, 1990) (Fig. 2). Species associations and fidelity in the various selected areas are examined with regard to factors such as their location (latitude, longitude), bottom depth, survey year (i.e. temporal component), and temperature changes (reflecting variability in Atlantic inflow). Qualitative and quantitative depictions relating species/community dominance to the major forcing functions are established and discussed within the context of predator–prey relationships and exclusion/preferenda concepts. Improved knowledge of regional fish communities, and the way they respond to environmental change, can contribute to more prudent management perspectives (Tyler *et al.*, 1982; Pimm and Hyman, 1989; Gabriel and Murawski, 1985; Mahon, 1985; Tonn, 1990).

Materials and methods

Trawl surveys

Research surveys, using a bottom trawl, have been systematically carried out by RV "Michael Sars" (Institute of Marine Research, Bergen) since 1980 at up to 24 "localities" in the Barents Sea for registering the deep-water prawn *Pandalus borealis* as well as fish by-catches. These "localities" are for the purposes of the present study further combined into eight larger "areas" ("North Finnmark", "Tiddly Bank", "Thor Iversen Bank", "East Bear Island Trough", "Hopen", "South Bear Island", "Southeast Bear Island" and "Southeast Garder Bank"), together forming a "compass"-shaped assembly with central, as well as north–south and east–west portions (Fig. 1). The total area studied can thus be considered to cover wide variations in latitude (ca. 70°N to 76°N), longitude (ca. 15°E to 35°E), and depth (ca. 150–500 m).

The trawl used is a "Campelen Super" 1800 mesh prawn-trawl in which the mesh size decreases from 80 mm in the anterior to 35 mm in the codend. The codend itself is lined with 8 mm mesh net, 4 m in length. The gear and its deployment have remained nearly unaltered since 1980 (Tavares and Øynes, 1980). The distance trawled in most hauls is three nautical miles (nm).

Although the main purpose of the surveys has been to monitor changes in the abundance and size composition of *P. borealis*, by-catches of the seven most economically important fish species (cod, *Gadus morhua*; haddock, *Melanogrammus aeglefinus*; redfish, *Sebastes* spp.; capelin, *Mallotus villosus*; polar cod, *Boreogadus saida*; Greenland halibut, *Reinhardtius hippoglossoides*; long rough dab, *Hippoglossoides platessoides*) have been recorded. From each haul the catch of *P. borealis* (kg, fresh wet weight) was recorded and a random sample taken for determination of length, sex, and reproductive stage. The lengths of the above fish species were also measured from the catches. The present study uses fundamental survey data from 1980–1989 (Tavares and Øynes, 1980; Teigsmark and Øynes, 1981, 1982, 1983; Hylen and Øynes, 1986; 1988; Hylen *et al.*, 1984,

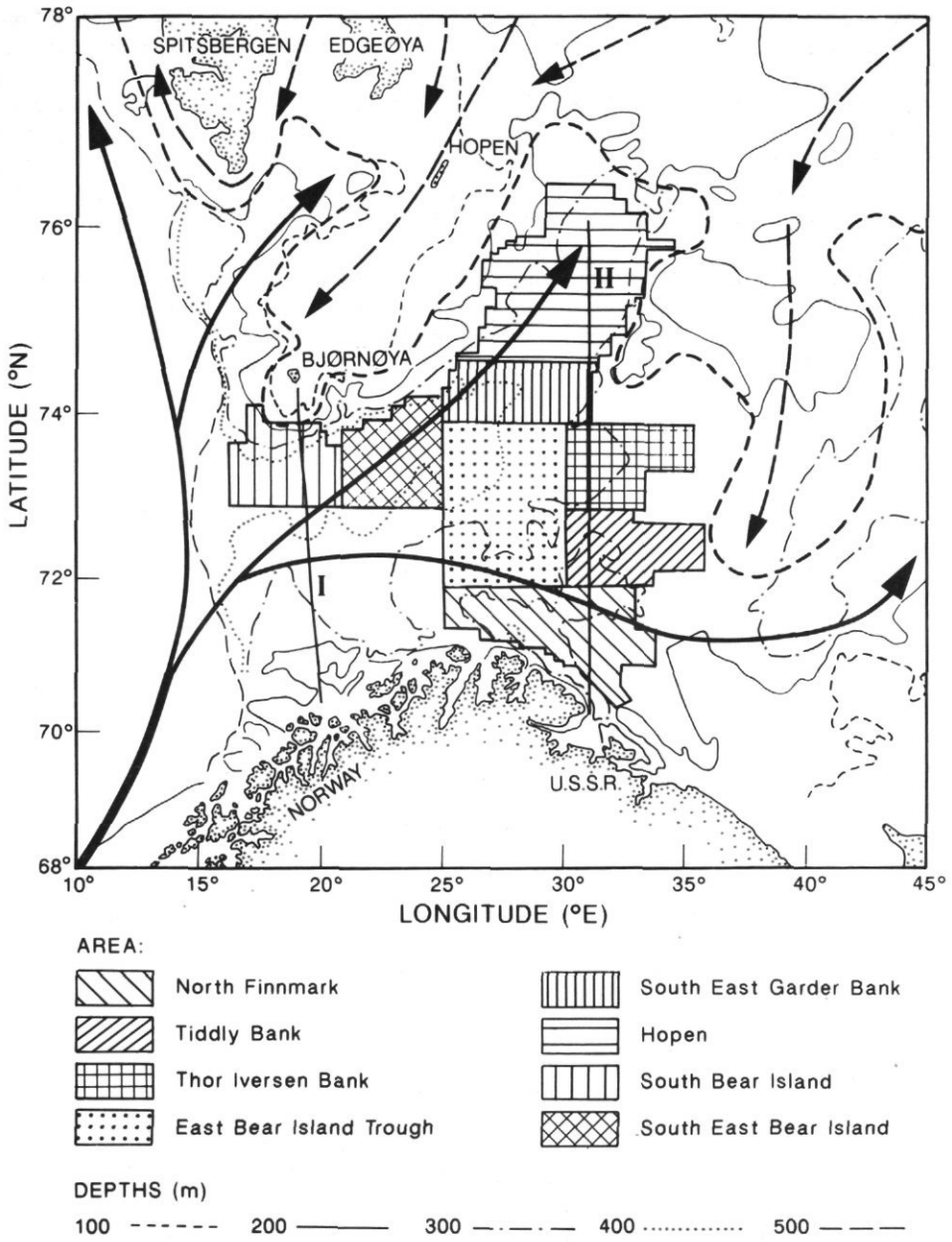

Figure 1. The study area of the Barents Sea for sampling of prawn–fish communities by bottom trawl in eight areas, designated "North Finnmark", "Tiddly Bank", "Thor Iversen Bank", "East Bear Island Trough", "Hopen", "South Bear Island", "Southeast Bear Island" and "Southeast Garder Bank". Details of topography (by 100 m contours), the classical position of the "Polar Front" (denoted by a dashed line), and major current branches (warmer, North Atlantic water = solid, northerly directed arrows; colder, Arctic water = dashed, southerly directed arrows). I = Fugløy–Bjørnøya hydrographic section, II = Vardø hydrographic section.

1987; Tveranger and Øynes, 1985; additional data kindly provided by IMR, Bergen), recording kg ("fresh" wet weight) of *P. borealis*, and numbers of fish by species (counts of individuals) per 3 nm towed distance. A total of 1072 trawl hauls were taken during this study. Bottom depths (m) were also recorded when the trawls were deployed.

Hydrographic data

Temperature data were recorded during the study period (1980–1989) by the Institute of Marine Research, Bergen, along hydrographic sections (I: Fugløy to Bjørnøya/Bear Island; II: Vardø northwards; see Fig. 1) running from the mainland northwards across the Barents Sea. Temperature anomalies were also used from

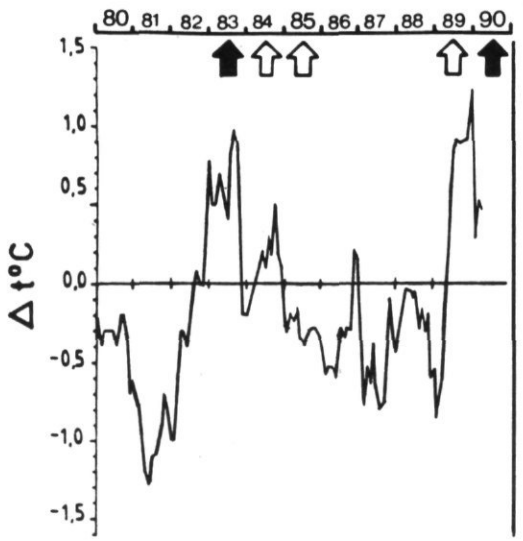

Figure 2. Temperature anomalies at the Kola hydrographic section (running northwards at 33°30'E) from 1980 to 1990. Dark arrows and light arrows = formation of high and medium abundance year classes, respectively, of cod. Temperature data from PINRO (Murmansk), after Loeng (1990).

the Kola section further east; the data from part of a continuous time series stretching back, with a break from 1940–1945, to 1920 (Sætersdal and Loeng, 1987; Loeng, 1989a, b; Midttun, 1989). The data are used to describe the gross temperature trends prevalent in the Barents Sea during the study period, and are also relevant for the matrix used in the multivariate analyses.

Additional temperature indices, presented as isotherms at 200 m depth from the Norwegian–Soviet 0-group fish surveys (August–September) of the Barents Sea (Anon., 1980–1989), were extracted annually for the original 24 localities, which together comprise the eight larger "areas". These data provide information on the position of the "Polar Front" and relevant transitions of temperature across the area of the trawl survey, as well as relevant data for the matrix used in the multivariate analyses.

Community, and species–environment relationships

Cluster analysis

Cluster analysis is a multivariate procedure commonly used for detecting natural "groupings" in ecological data (Gauch and Whittaker, 1981; Field *et al.*, 1982; Gauch, 1982; Pielou, 1984). In the present study we define the *community* as an assemblage of species populations which occur together in space and time (Begon *et al.*, 1986), and use cluster analysis to appraise how many species groups were present (eight areas combined) as well as the annual stability of particular species within these groups during the study period. Matrices for clustering by Bray-Curtis and Pearson product-moment correlation were constructed using the SIMINT program, while the clustering computations were performed using the SAHN program of NTSYS-pc (Rohlf, 1989).

Canonical correspondence analysis

A common problem in community ecology is discovering how a multitude of species respond to external factors such as environmental variables. Regression and ordination have recently been integrated into techniques of *multivariate* direct gradient analysis, called canonical ordination (Jongman *et al.*, 1987; ter Braak and Prentice, 1988). The Fortran program CANOCO (*cano*nical *c*ommunity *o*rdination, version 3.10) includes canonical correspondence analysis (CCA), which escapes assumptions of linearity between species abundance and gradients (cf. Principal Components Analysis) and is able to detect unimodal relationships between species and "environmental" variables (ter Braak 1986, 1987a, b).

The CCA option of CANOCO was used to examine the relationship between the abundance of the eight species entities trawled in the Barents Sea, and to relate the abundance/dominance of given species to the eight selected geographical areas (localities). The inclusion of "environmental factors" (bottom depth, latitude, longitude, water temperature) and year of sampling allows the degree of stability and characteristics of the species and their "domains" to be evaluated. Presentation of an ordination diagram, with inclusion of species, sampling localities (by year) and environmental factors, allows one to search for relationships: species which are arrayed close together have similar distributions, and stations/localities which are arrayed close together have similar species attributes. Species in close proximity to particular stations/localities will be more "dominant" (i.e. over-represented) than those further removed. The major points to be borne in mind when interpreting the CCA plot in CANOCO are (ter Braak, 1986, 1988):

(1) environmental factors with long "arrows" are better correlated with the axes than those with shorter "arrows", and accordingly more strongly related to the species pattern in the plot;
(2) "arrows" point in the direction of maximal change of given environmental factors in the plot, and the length of the arrow is proportional with the degree of change (i.e. larger change reflected by long "arrows");
(3) species can be projected relative to the "arrows", such that the ordering of species along the axis of the "arrow" is approximately the ranked, weighted median value of the species relative to the environmental factor;

(4) the absolute length of the "arrow" is immaterial; it is the relationship between its length and direction that is important;
(5) the proximity of species plotted relative to localities denotes the degree of influence/dominance of the species at the localities.

Forward selection of environmental variables

CANOCO was also used to select, by forward (stepwise) selection of environmental variables, a minimum set of investigated environmental variables that explain the variation in the species data. At each step CANOCO can test whether the additional effects of a particular variable on the species are statistically significant. This is achieved by means of a Monte Carlo permutation test, which replaces the F-test or *t*-test in forward selection in univariate multiple regression (ter Braak, 1990).

Results

Temperature variations in the Barents Sea, 1980–1989

The inflow of Atlantic water into the Barents Sea during the study period (1980–1989) exhibited great variability as measured in the form of temperature anomalies (Δt °C) at the Kola section (Fig. 2, after Loeng, 1990). The period started with slightly below normal conditions, falling to a minimum in 1981 before rising towards a peak in 1983, followed by a general downward trend (incorporating several transient rises) reaching a low in late 1988 before steeply intensifying to a maximum in 1989. The trough in 1981 was one of the coldest recorded, while the peaks in 1983 and 1989 were among the warmest recorded since records started in the 1930s.

The "near-bottom" (i.e. within 100 m of the bottom) waters along a south to north (70.5°N–76.5°N) section of the Barents Sea are clearly affected by different temperature regimes (Fig. 3A–H). The southern area (<ca. 73.5°N) is generally covered by "warmer" waters in excess of 2°C, while the northern area (>ca. 73.5°N) is generally covered by significantly "colder" waters (<ca. 1°C). However, annual variability is apparent and several gross situations may be registered:

Under high Atlantic Influence the southern part of the

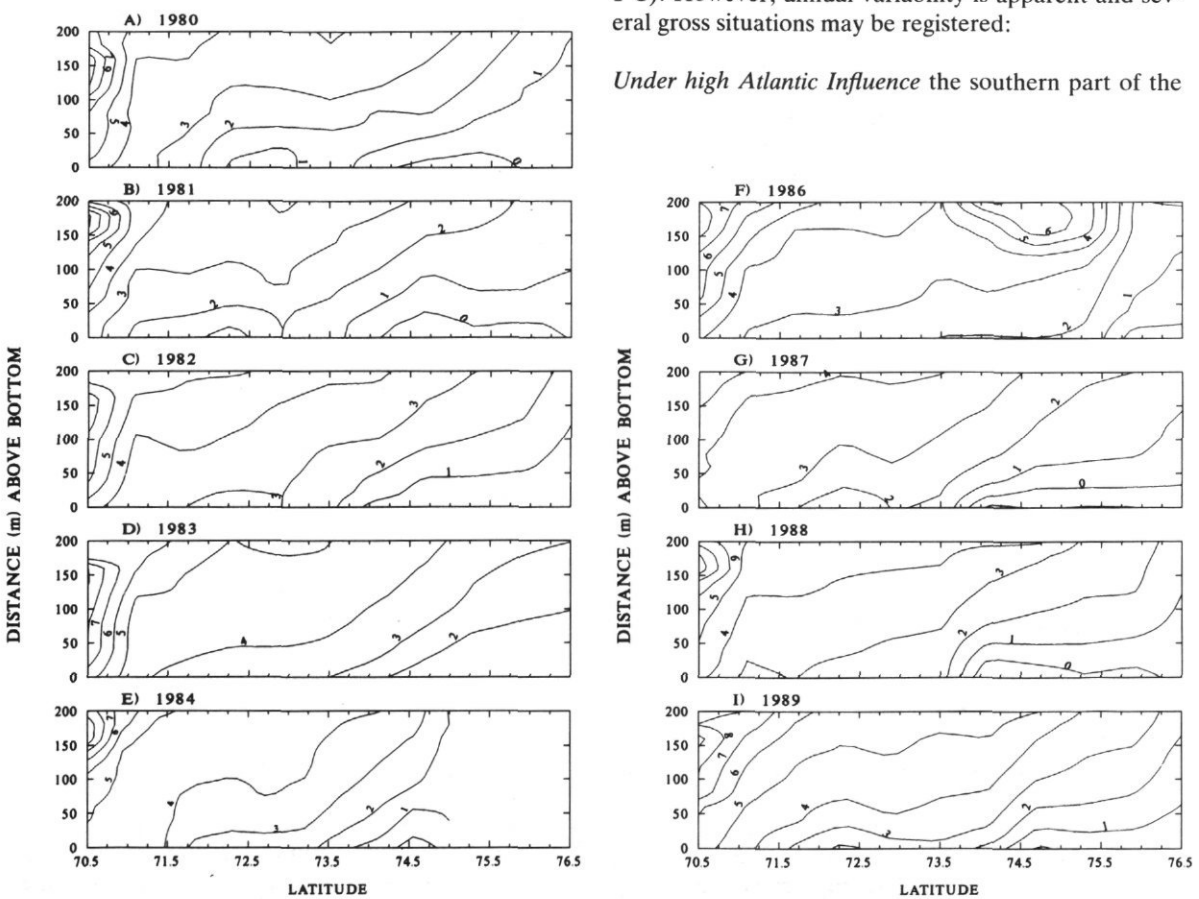

Figure 3. Temperature isopleths (°C) related to latitude and depth (m) above the bottom along the Vardø section of the Barents Sea in (A) 1980, (B) 1981, (C) 1982, (D) 1983, (E) 1984, (F) 1986, (G) 1987, (H) 1988, (I) 1989.

Barents Sea exhibits conspicuously elevated near-bottom temperatures, while the colder waters of the northern area are displaced far northwards. Good illustrations of this are seen in the isotherm distributions of 1983 and 1989 (Figs. 3D and I, respectively).

Under low Atlantic influence the southern part of the Barents Sea exhibits decreased near-bottom temperatures as the warmer waters are displaced towards the south, and colder waters (<ca. 1°C) encroach over the northern region. Good illustrations of this are seen in the isotherm distributions of 1980 and 1981 (Figs. 3A and B, respectively).

Variations in abundance of fish and prawn, 1980–1989

Most of the species examined in this study exhibited an approximately 10-fold range in their abundance during the study period (Figs. 4A and B). Prawns proved an exception, probably partly due to their abundance being measured in biomass rather than as individuals. At the start of the study (1980–1982), polar cod and capelin exhibited high abundances while cod and haddock exhibited the reverse (Fig. 4A). A particularly strong year class of cod was produced in 1983, followed by moderately strong year classes in 1984 and 1985 (see also Fig. 2). Variations in abundance of cod and haddock showed essentially similar trends throughout the study period, but after 1985 haddock abundances declined while those of cod remained basically high. After 1982, both polar cod and capelin were generally less abundant. Redfish were by far the most abundant species, and showed a sustained increase between 1980 and 1984 before numbers declined again (Fig. 4B). Greenland halibut exhibited a basic increase in abundance during the study period, but temporary declines occurred in 1983 and 1989. Long rough dab and "others" (several less abundant fish species, including flatfish) were second in abundance to redfish.

Examination of species abundances in the eight areas studied (Figs. 5 and 6) indicates that cod and haddock on the one hand and polar cod on the other have clear differences in their geographical distribution of abundance. Cod were generally most abundant in the southern and central areas (North Finnmark, East Bear Island Trough, South Bear Island, Tiddly Bank, and Thor Iversen Bank), while they were virtually absent (with only a weak infringement of the area associated with the abundant year classes produced in the middle of the study period) from the Hopen area (Fig. 5). Polar cod were most abundant in the northern, Hopen area throughout the study and least abundant in the southern-central, North Finnmark and East Bear Island Trough areas (Fig. 6). Abundances of polar cod clearly varied most in the eastern, Thor Iversen Bank, and

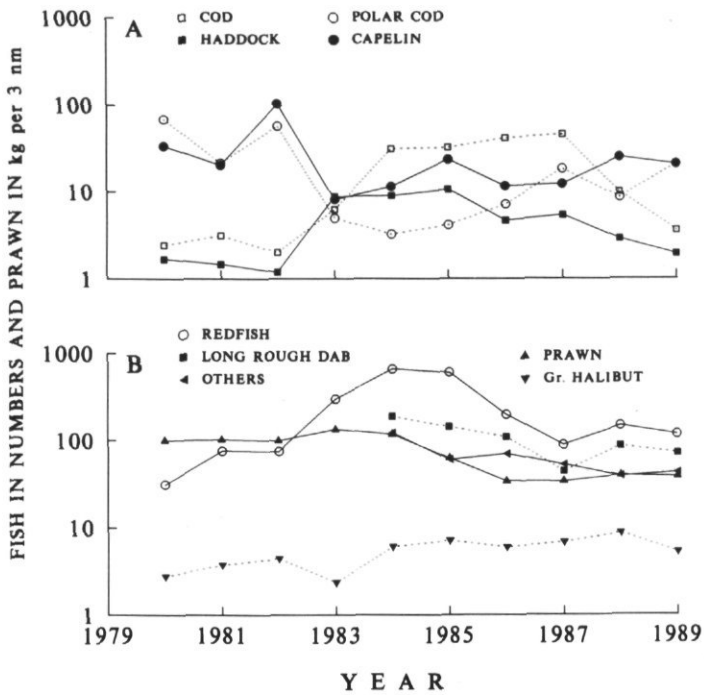

Figure 4. Variations in the abundance (numbers per 3 nm trawled bottom distance) of fish and prawn (*Pandalus borealis* in kg per 3 nm) caught in demersal trawl surveys (1980–1989) of the Barents Sea (eight areas combined). (A) Cod, haddock, polar cod, capelin, (B) prawn, redfish, Greenland halibut, long rough dab, "others".

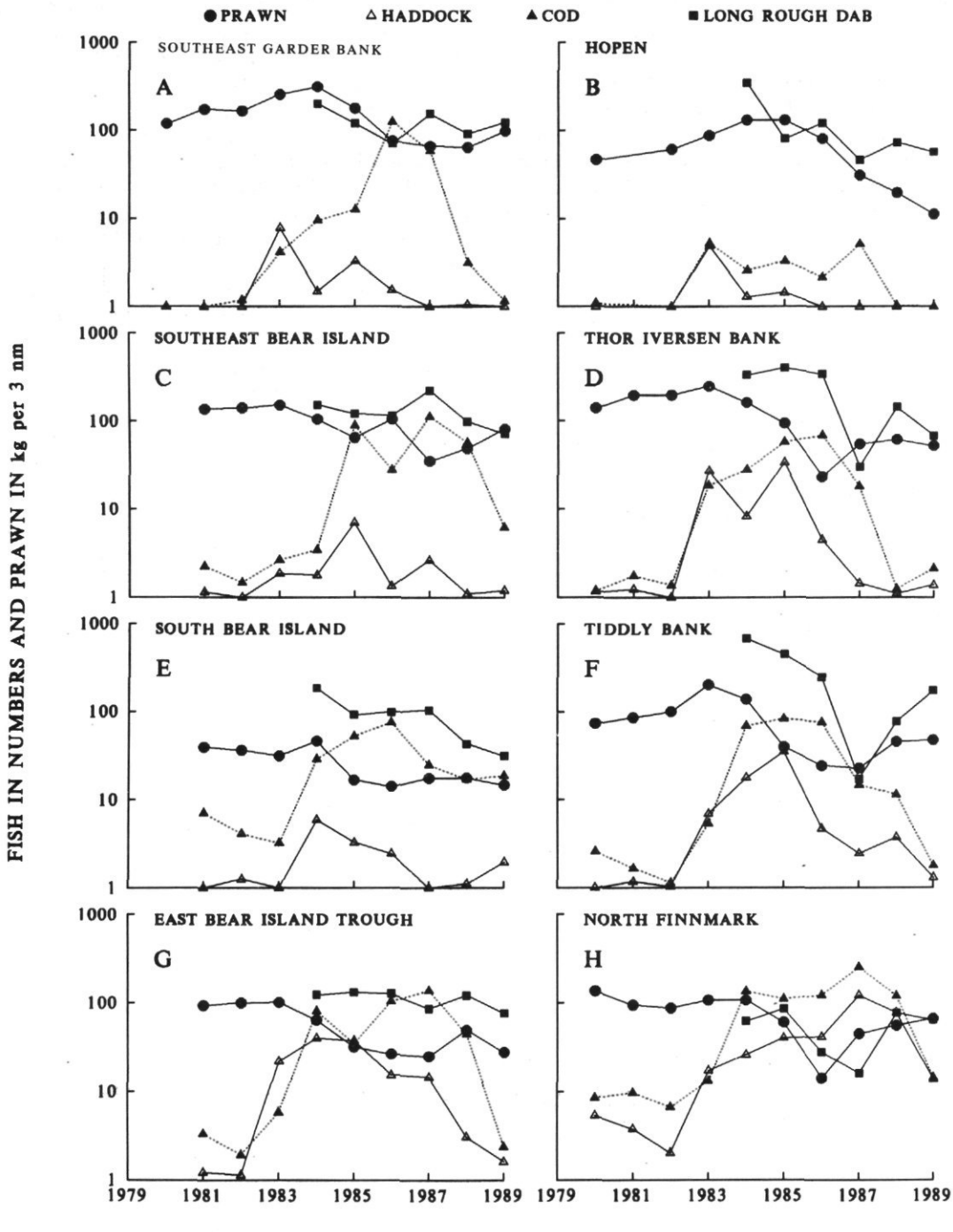

Figure 5. Variations in the abundance (numbers per nm trawled bottom distance) of cod, haddock, long rough dab, and prawn (*Pandalus borealis*) caught in demersal trawl surveys in eight areas (A–H) of the Barents Sea from 1980 to 1989.

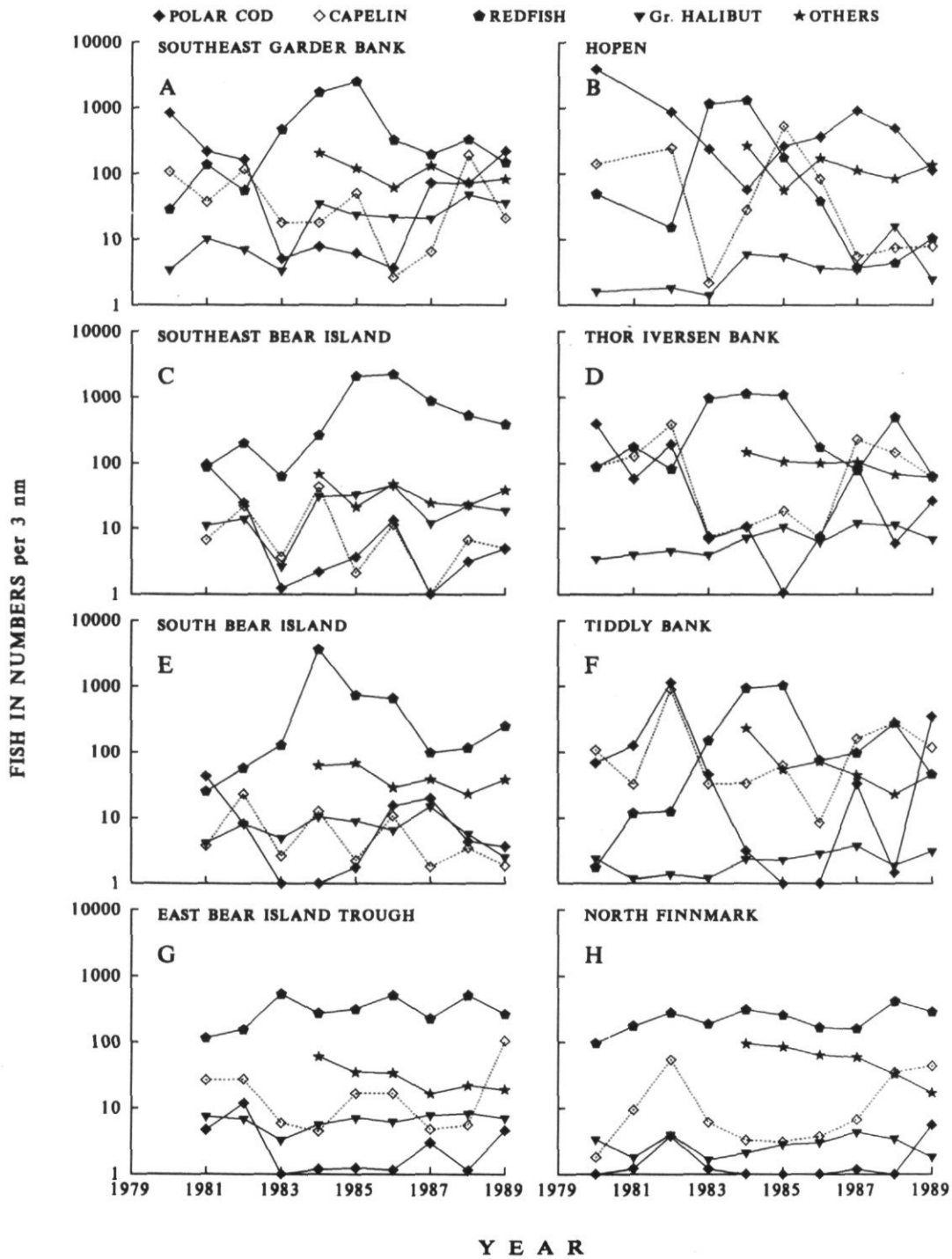

Figure 6. Variations in the abundance (numbers per 3 nm trawled bottom distance) of polar cod, capelin, redfish, Greenland halibut, and "others" caught in demersal trawl surveys in eight areas (A–H) of the Barents Sea from 1980 to 1989.

Tiddly Bank areas. Capelin were moderately abundant in all areas, but were more numerous in the Hopen, Southeast Garder Bank, Thor Iversen Bank and Tiddly Bank areas. Prawns were generally highly abundant throughout all the eight areas during the study period, but clear declines in their abundances were apparent in a number of areas (e.g. Tiddly Bank, Thor Iversen Bank) as cod, one of their major predators, became more abundant. Although redfish were abundant in all eight areas there were signs of inverse relationships in their abundances with those of polar cod in the Hopen, Tiddly and Thor Iversen Banks, and South Bear Island areas.

Demersal prawn–fish communities

Basic species relationships

Bray-Curtis dendrograms (Fig. 7) demonstrate the degree of similarity between the species distributions and their densities and the annual variability, while correlation (Pearson product-moment) dendrograms (Fig. 8) demonstrate the manner in which particular species abundances are positively or negatively related, irrespective of densities, during the study period. Irrespective of clustering technique, it is evident that cod and haddock abundances/distributions increase and decrease together (see also Figs. 4A and B and 5 and 6), as do those of polar cod and capelin. Overall, the relationship between polar cod and capelin is less distinct than that of cod and haddock. In some years capelin becomes less closely attached to polar cod in the dendrograms; this is particularly apparent in the exceptionally warm year of 1983. A greater variability in "fidelity" exists in the remaining species, which are relatively intermediate in the classification.

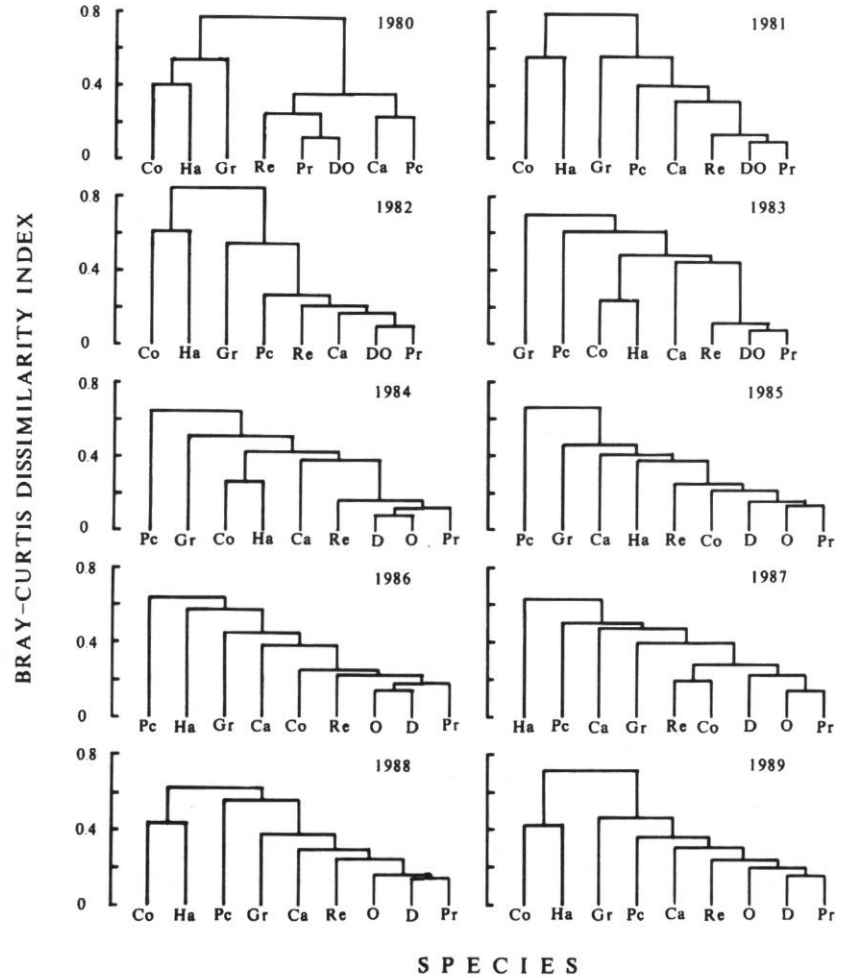

Figure 7. Bray-Curtis similarity dendrograms of fish and prawn (*Ha*ddock, *Co*d, *Re*dfish, *Gr*eenland halibut, long rough *D*ab, "Others", *Ca*pelin, *P*olar cod, and *Pr*awn. Prior to 1984 *D*abs and *O*thers formed a single entity, *DO*) caught in demersal trawl surveys of the Barents Sea (eight areas combined) between 1980 and 1989.

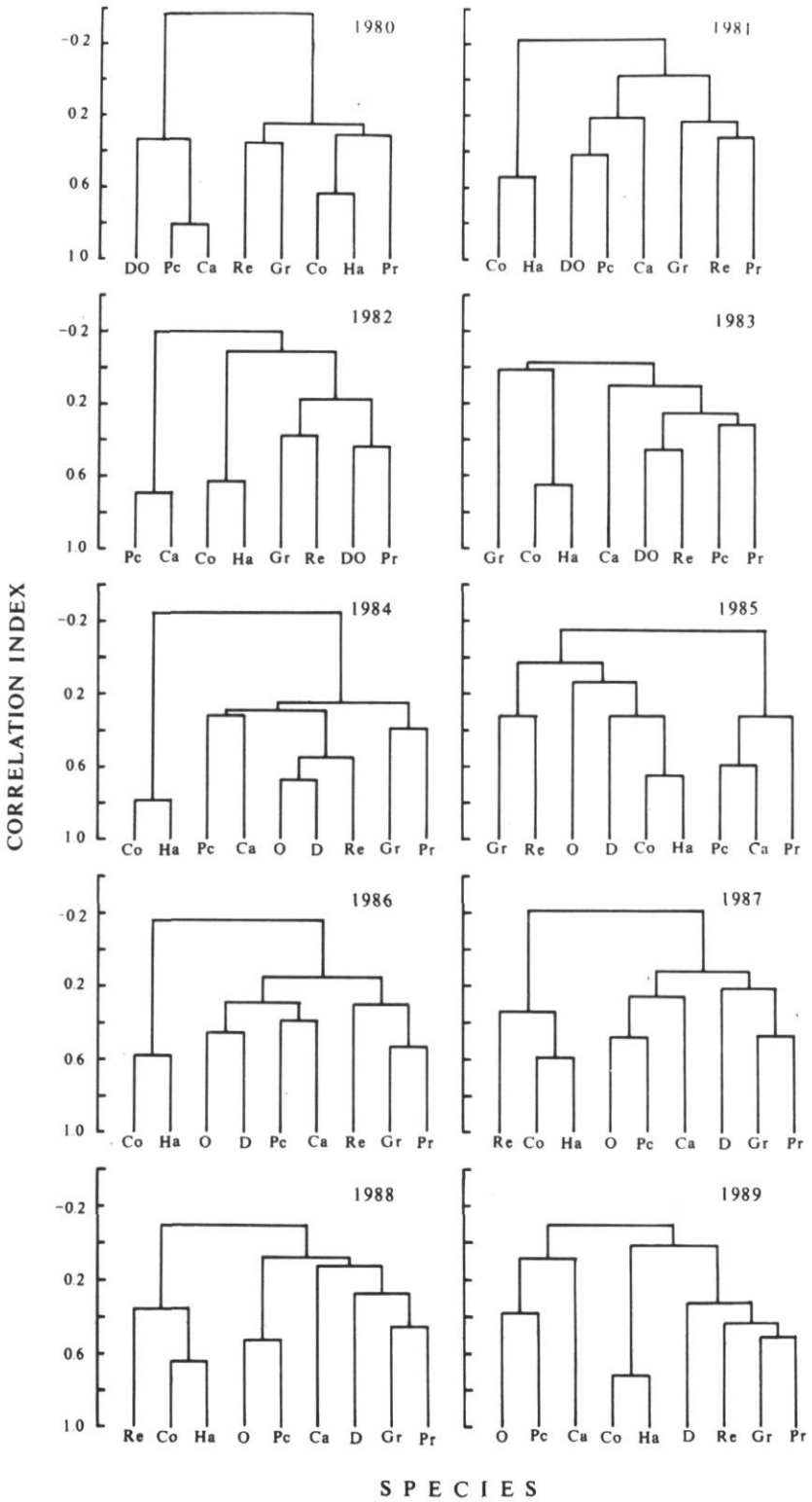

Figure 8. Correlation (Pearson product-moment) dendrograms of fish and prawn (see legend to Fig. 7 for species key) caught in demersal trawl surveys of the Barents Sea (eight areas combined) between 1980 and 1989.

Species–environment relationships

Canonical correspondence analysis (CCA) plots of the first two axes for the species–environment data (using sampling year as covariate) are presented in Figure 9. The first two axes accounted for 95.5% (Axis 1 = 73.6%, Axis 2 = 21.9%; ratio of 3.4:1) of the variance of the species–environment relationship. The species–environment correlations were 0.68 and 0.63 for Axis 1 and Axis 2 respectively. Latitude and temperature were negatively correlated and the major environmental determinants of Axis 1, while longitude and depth were negatively correlated and the major determinants of Axis 2 (Fig. 9 and Table 1). It is also evident that temperature is partially correlated with both latitude and longitude, indicating that the southwestern area is particularly warm while the northeast area is cold.

The species–environment relationship indicates that haddock and cod abundances in particular were positively correlated with higher temperature and low latitude. Polar cod, and capelin to a lesser extent, were positively correlated with high latitude and low temperature. Abundances of redfish and long rough dab, as well as biomass of prawns, are distributed in a relatively transitional position along the temperature gradient, indicating that these species are relatively eurythermal compared with more stenothermal species like haddock, cod and polar cod. Abundances of Greenland halibut are positively correlated with greater depths.

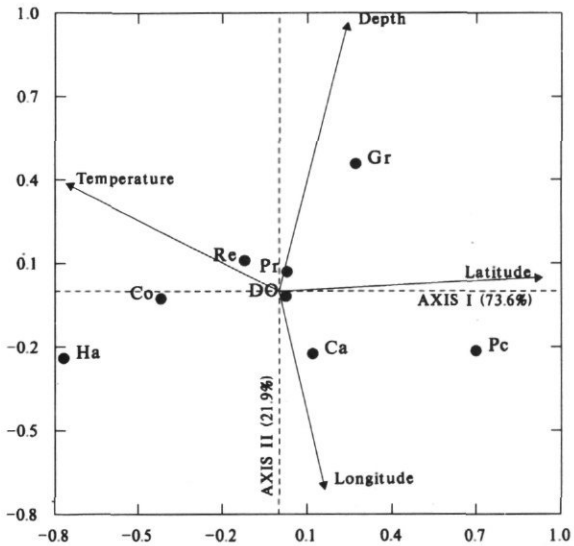

Figure 9. Canonical correspondence analysis (CCA) ordination plot of Axis I and Axis II, relating variations in the abundance of species (*Cod*, *Haddock*, *Redfish*, *Polar cod*, *Capelin*, *Greenland halibut*, "long rough *Dab* + *Others*", and *Prawn*) caught in demersal trawl surveys in the Barents Sea to environment (temperature, depth, latitude, and longitude) from 1981 to 1989.

Table 1. Weighted correlation matrix (weight = sample total) produced from canonical correspondence analysis of species–environment relations from the Barents Sea, 1981–1989.

	SpAx1	SpAx2	EnAx1	EnAx2	Lat	Long	Temp	Depth
SpAx2	-0.20	1.00	0.00					
EnvAx1	0.68	0.00	1.00					
EnvAx2	0.00	0.63	0.00					
Lat	0.62	0.03	0.92	0.05	1.00			
Long	0.11	-0.44	0.16	-0.69	-0.09	1.00		
Temp	-0.52	0.25	-0.76	0.39	-0.58	-0.39	1.00	
Depth	0.16	0.61	0.24	0.97	0.27	-0.66	0.17	1.00

Geographic and annual variability: species–environment influences

CCA may be used to examine the annual variability in the influence of *species* and/or *temperature* affecting the eight geographical areas. This is achieved by plotting and identifying the areas by year in the species–environment data space (Fig. 10). The data matrix comprises information from 998 out of a total of 1072 demersal trawl-hauls taken during the study period.

Although there is a large amount of annual variation, areas representing the extremities of the compass-rose are well separated by various aspects of the species–environment gradient. For example, the Hopen (northern) and the North Finnmark (southern) areas are clearly separated. The South Bear Island and Southeast Bear Island areas are similar, exhibiting relatively homogeneous environment and species characteristics.

The North Finnmark (southernmost) area is obviously primarily characterized by the dominance of haddock and cod and warmer conditions. The sequence 1983 to 1984 is indicative of warming corresponding with good recruitment of cod and haddock, while mainly decreasing temperatures and decreased dominance of cod and haddock (partly related to poor recruitment) lasts from 1986 to 1989.

Although the Hopen (northernmost) area is obviously primarily characterized by its dominance by polar cod and cold water, a marked influence of warmer water and the presence of cod and haddock in the area is clearly seen in the sequence 1982 to 1984. The sequence 1984 to 1987 is clearly influenced by decreasing temperatures and enhanced dominance by polar cod. Although 1989 was a particularly warm year based upon temperature anomalies (Fig. 2), relatively cold waters still covered the bottom in the Hopen area (Fig. 3I) that year.

The Tiddly Bank area, in the east, showed great variability during the period, with the major progression moving along the temperature and cod/haddock–polar cod axis. The sequence from 1982 to 1984 is considerable and clearly directed towards warming and increased dominance by cod and haddock, while the large move-

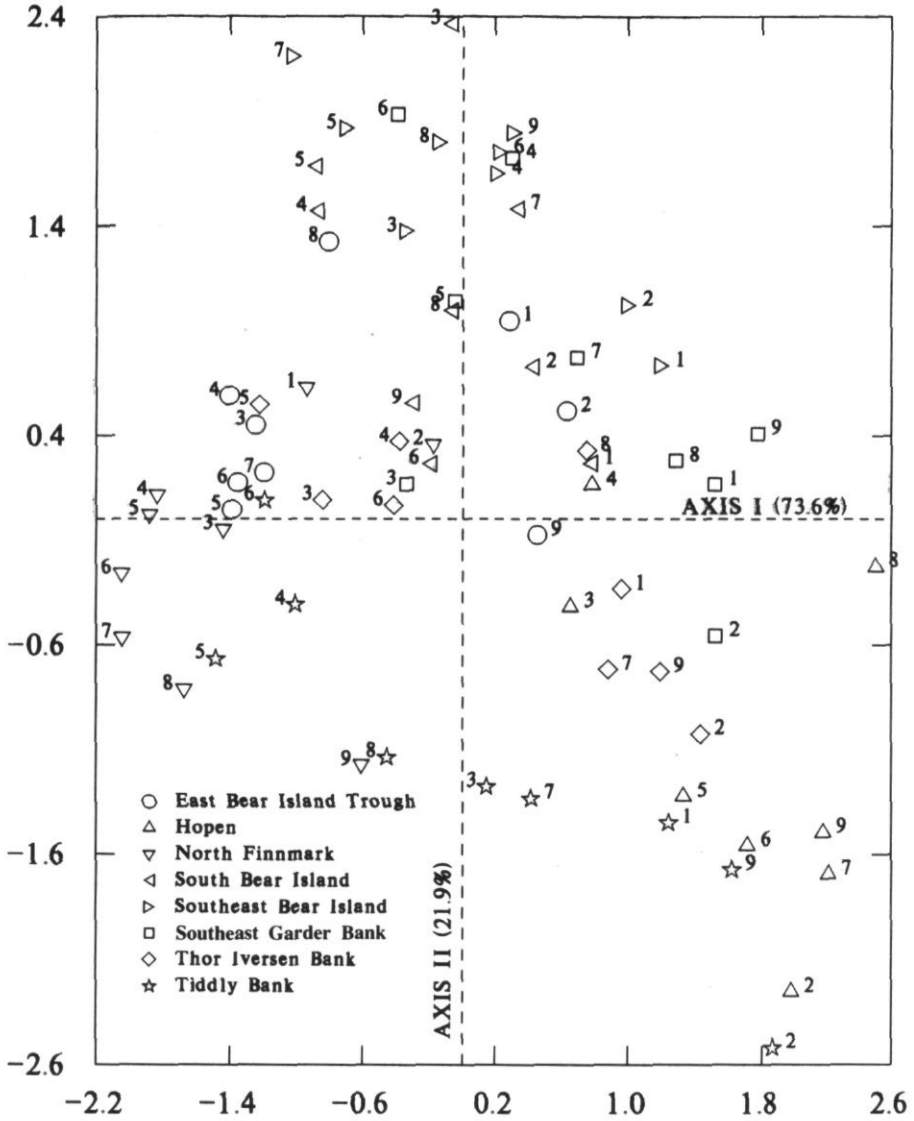

Figure 10. CCA ordination plot showing variation in the location of the eight sampling areas of the Barents Sea within the species–environment data space of Fig. 9 from 1981 to 1989 (years denoted by numerals 1–9).

ment seen from 1986 to 1987 reflects a clear reversal in temperature with increased dominance by polar cod. The Tiddly Bank is close to the southeasterly extension of the Polar Front (see Fig. 1).

The Thor Iversen Bank area, slightly further away from the Polar Front than the Tiddly Bank, is also a transitional area regarding species dominance and temperature regime. The sequence from 1982 to 1983 is clearly directed towards warming and increased influence by cod and haddock, whereas the sequence from 1985 to 1987 reverses towards colder water and polar cod influenced conditions. The warmer regime/community at the Thor Iversen Bank area resembles the North Finnmark area under a colder regime/community, while the colder regime/community at the Thor Iversen Bank overlaps with the Hopen area when it has a warmer regime/community.

The East Bear Island Trough (southeast) area, besides being associated with greater depth, is characterized by generally warm conditions and a wide range of species (e.g. redfish, Greenland halibut, long rough dab, cod, and haddock) in the community. Polar cod contribute little in abundance. Clear changes in species dominance are difficult to discern. The most evident temperature-related trends (i.e. progressions along the temperature axis) are the cold sequences from 1981 to 1982 and from 1988 to 1989.

Table 2 provides a summary of the clearly delineated

Table 2. Summary of periods characterized by "elevated" (*W*arm or increasing) and "reduced" (*C*old or decreasing) bottom temperatures in selected areas of the Barents Sea as clearly illustrated "sequences" from CCA ordination of Figure 10.

Hopen	(W: 1982–1984; C: 1984–1987)
SE Garder Bank	(W: 1982–1983; C: 1986–1987)
Thor Iversen Bank	(W: 1982–1983; C: 1985–1987)
Tiddly Bank	(W: 1982–1984; C: 1986–1987)
E. Bear Island Trough	(W: 1981–1982, 1988–1989)
North Finnmark	(W: 1983–1984; C: 1981–1982, 1986–1989)

"warm" and "cold" temperature trends extracted from the CCA ordination (Fig. 10).

Forward selection of environmental variables

Of the 10 years examined, 50% of the primary factors (*first step*) explaining species distributions were temperature, while 50% were latitude (Table 3). However, the importance of temperature is emphasized by the ranking of the second position occupied by environmental variables at the first step; temperature never falls below the second rank, while latitude exhibits a relatively variable contribution. At the *second step*, after temperature or latitude were incorporated at the first step, depth was clearly denoted as the second most prominent environmental explanation. At the *third step*, longitude accounted for the majority (70%) of the remaining factors for inclusion. At the *fourth step*, where only one remaining factor was available, either temperature or latitude accounted for 80% of the final factors for inclusion. The situation at the fourth step, when compared with the first step, reflects which of these two environmental factors was first included.

Temperature was the dominant primary explanatory environmental variable during the 1982 to 1983 period of "warming", while latitude was dominant during the "cold" period of generally decreasing temperatures from 1984 until 1987 (see Fig. 2).

Discussion

Early estimates of the volume flux of the Norwegian Atlantic current were about 3 Sv (1 Sv = 10^6 m^3 s^{-1}) (Helland-Hansen and Nansen, 1909), but contemporary research suggests that this is more likely to be in the order of 8 Sv on average (Worthington, 1970). Annual and monthly variations may be great, probably ranging from 5 to 15 Sv (Gould *et al.*, 1985; McClimans and Nilsen, 1990). Although the volume flux of Atlantic water into the Barents Sea is obviously much less than this, the variability is likely to be appreciable. A recently developed, wind-driven numerical current model suggests concord between variability in the volume in-

Table 3. Forward selection of environmental variables (in 4 steps, ST1-4), using Monte Carlo perturbation analyses, explaining the investigated species distributions in the Barents Sea by year (1980–1989). Variables: temperature, Latitude *N*, Longitude *E*, *D*epth. Numerals represent F-ratio.

	<----------YEAR---------->									
	80	81	82	83	84	85	86	87	88	89
ST 1	**T42.1** E31.2 D13.8 N 6.0	**T22.0** E20.2 N10.7 D 9.7	**T35.6** E18.1 N17.7 D15.1	**T31.2** E19.1 D13.7 N 9.9	**N50.4** T33.5 D20.2 E 5.2	**N71.2** T58.3 E29.9 D20.5	**N53.5** T37.0 D16.1 E13.2	**T60.1** N32.8 E30.4 D 7.8	**N49.4** T30.8 E18.4 D 8.0	**N38.2** T34.7 E16.2 D14.1
ST 2	**N 4.1** E 2.2 D 1.0	**D12.8** E 8.2 N 6.7	**D20.5** N11.7 E 6.5	**D14.3** E 8.0 N 6.8	**D16.2** E11.0 T 5.4	**D31.3** E16.5 T 5.1	**D23.5** E15.1 T 3.7	**D12.0** N 6.1 E 4.2	**E15.5** D11.8 T11.4	**E24.8** D16.7 T16.2
ST 3	**E 2.4** D 1.1	**E12.1** N 5.1	**E10.9** N 3.5	**N 2.0** E 1.7	**E18.1** T 8.2	**E 6.5** T 5.5	**E 5.5** T 3.6	**E 4.4** N 4.0	**D 7.1** T 5.7	**D19.5** T 4.2
ST 4	D 1.8	N 1.5	N 7.4	E 3.2	T 2.0	T 4.8	T 2.3	N 4.8	T 6.9	T 3.2

flux of Atlantic water and temperature anomalies in the Kola section (Ådlandsvik, 1989). Temperature anomalies at the Kola section and depth–latitude temperature profiles clearly demonstrate a high degree of hydroclimatic variability between 1980 and 1989.

Although the location of the Polar Front in the Barents Sea is typically presented as in Figure 1 (cf. Loeng, 1989a, b) the present study indicates that this is more applicable to the surface waters than the near-bottom ones. Cold (<1°C) and dense bottom water, formed by winter cooling of the surface waters, may some years extend over large areas of the bottom to as far south as 73.5°N of latitude along the eastern Vardø Section of the study area. Temperature anomalies are obviously useful indices of gross temperature trends, but when they integrate temperatures over depth and over latitude (e.g. Fig. 2) they may provide misleading depictions of temperature conditions for demersal communities in particular regions. Although, Figure 2 suggests that 1987 was appreciably colder than 1988, and despite this being supported in the depth–latitude sections (Fig. 3) in the southern half, the demersally appropriate temperatures north of 73.5°N are similar in both years. Again, for example, integrated depth–latitude anomalies suggest similarity between the exceptionally warm years of 1983 and 1989 (Fig. 2), whereas there are very clear differences in the northern demersal-regime in these years. Annual variation in the distribution of the cold bottom-water in the Barents Sea will probably exhibit a time-lag relative to the cycles of the surface watermasses (see Midttun, 1985, 1989).

Variability in the influx of Atlantic waters to the Barents Sea plays an important role in influencing the abundance and distribution of species of fish and benthos (Loeng, 1989a, b; Blacker, 1965). The majority of the year classes of high and medium abundance of the Northeast Arctic cod are either associated directly with positive temperature anomalies in the early part of a warm period in the Barents Sea, when feeding areas are expanding, or they occur immediately prior to a shift to a warmer regime (Sætersdal and Loeng, 1987; Loeng, 1989a). Year classes of low abundance are common, occurring in about 66% of all years, and are equally likely to occur in both cold and warm years. Medium strength year classes occur twice as frequently in warm years than they do in cold ones, while strong year classes are 12 times more frequent in warmer than colder years. The influence of temperature is further emphasized when one considers that on average a strong year class is about 10 times more abundant than a weak year class and about twice as abundant as a medium one (Sætersdal and Loeng, 1987).

The present study, using both cluster analysis and canonical correspondence analysis (CCA), clearly demonstrates the high degree of fidelity between cod and haddock on the one hand and between polar cod and capelin on the other. Concurrent fluctuations in year-class strength and recruitment of cod and haddock have been described for northwest Atlantic (Koslow, 1984) and Icelandic (Pálsson, 1984) stocks. According to Ekman (1967), most of the fish species of the Barents Sea fall into three zoogeographical groups: arctic, arctic boreal, or boreal, although some warm-water boreal species may also be found. Relative to this classification, the present study clearly establishes Barents Sea cod and haddock in the boreo-Atlantic domain, polar cod in the arctic domain, with capelin occupying a transitional, "arctic boreal" position. CCA and the cluster analysis demonstrate that capelin is not an arctic species *sensu stricto*, in contrast to a tendency for it to be classified as an arctic species (cf. Ekman, 1967; Dragesund and Gjøsæter, 1988). However, our findings emphasize (on the basis of species–environment ordination and classification) that the individualistic community concept, where *discrete* community boundaries are rare (see Begon *et al.*, 1986) is essentially prevalent in the Barents Sea: a given species which occurs in one predictable association is also likely to occur with another group of species under different conditions elsewhere.

Variability in the abundance and dominance of these four key species were clearly highly correlated with a marked temperature gradient as shown by CA; cod and haddock abundances and distributions were positively correlated with warmer conditions, while polar cod, in particular, and capelin, to a lesser extent, were positively correlated with colder conditions. Temperature showed a significant degree of covariation with latitude and longitude (both with inverse correlations), such that the areas showing the greatest degree of difference in species–environment gradient represented the extremes of the compass rose (Zenkevitch, 1963; Loeng, 1989a, b). As expected, the northern Hopen area was markedly affected by periods of warm water advection (e.g. 1983). Nevertheless, the eastern transitional areas (Tiddly and Thor Iversen Banks) were also greatly affected by surges of warm and cold water. This emphasizes that high latitude "frontal seas", such as the Barents Sea, are likely to be especially affected by global climate change (Øiestad, 1990).

The present study ranged over about 6° of latitude and 20° of longitude (similar in area to the North Sea), depths of 100 to 500 m, and temperatures from about 0°C to 6°C. The CCA and forward selection of environmental variables explaining variations in species abundance indicated that temperature and latitude were equally frequent *primary* explanatory variables (accounting for 50% each) during this investigation, while depth was the most frequent environmental variable to be included at the second step, and longitude was the major explanatory variable at the third step (Table 3). Although depth and temperature are frequently identified as major factors delineating demersal fish

assemblages, latitudinal and longitudinal influences are rarely detected due to the generally limited geographical extent of individual studies (Gomes *et al.*, 1989).

The axis of the temperature gradient was also partly correlated with latitude and longitude, in accordance with the well-known trend in the Barents Sea for temperature to decrease from north to south and from west to east (Dragesund and Gjøsæter, 1988; Wielgolaski, 1990). The results established here for cod clearly support the previously mentioned proposals regarding the high probability of strong year-class formation when appreciable warming occurs. Furthermore, our results also underline that elevated inflow of warmer water into the Barents Sea will be beneficial to the young cod in distributing them over a larger warmer domain, and so contributing towards reduced likelihood for intraspecific competition than would otherwise occur. Sætersdal and Loeng (1987) have similarly noted that the favourable feeding areas for cod larvae in the eastern Barents Sea will be expanded in warm years. It is generally accepted that when the temperature is high cod are distributed easterly in the Barents Sea, while westward movement occurs with low temperatures (Hylen *et al.*, 1961; Midttun *et al.*, 1981), and this results in more westerly deployment of fishing in cold years than in warm years (Eggvin, 1938). When temperatures in the Barents Sea fall (i.e. negative anomaly increases) the older cod move westwards (Midttun *et al.*, 1981); thus an extreme westerly distribution of the biomass of the cod is characteristic of low temperatures and relatively old ages (4–7) of the dominant year classes. This is in accordance with the observations and the CANOCO predictions towards the end of the decade with poor recruitment and declining temperatures.

Adult capelin feed pelagically in the northern areas of the Barents Sea during the summer and autumn, while younger individuals have a more southerly distribution (Loeng, 1989a; Falk-Petersen *et al.*, 1990). Adult capelin may be found in both Atlantic and Arctic waters, generally within colder waters (Loeng, 1989b; Wielgolaski, 1990). Loeng (1989a) concluded that although capelin seem to respond to a relatively slow decrease in temperature from 1976 to 1979 by moving southwestwards, the mean concentrations of two to four-year-olds were found in a constant position relative to the temperature field. Our CCA ordination clearly reveals that the abundance and distribution of capelin from 1980 to 1989 were correlated with decreasing temperature, but to a much lesser extent than polar cod. Within the species–environment relationships seen here, it is highly likely that capelin will thrive in *the more transitional* regimes between "warmer" and "colder" periods.

Given the results of the present study, it is tempting to comment on the events that have occurred in the Barents Sea leading to the "crash" of the capelin stocks in 1983/1984 and clear signs of depressed growth rates exhibited by the 1983 year class of cod (Mehl, 1989; Hopkins and Nilssen, 1991). Prior to 1983, the last medium strength year class of cod was produced in 1975 and the previous strong year class in 1970; thus emphasizing that only weak recruitment of cod occurred between 1975 and 1983 (Loeng, 1989a). This period of weak cod recruitment coincided with a period of cooling (reduced influx of Atlantic water), reducing the dominance of the "cod-haddock" regime. The capelin fishery developed as a replacement for the herring (whose spawning stock and recruitment had collapsed by the early 1970s) during this period of cooling, peaking in 1977 and catching close to two million tonnes until its marked decline in 1983 (Hopkins and Nilssen, 1991); 1982/1983 represented the first major positive anomaly since the mid-1970s, and would accordingly have favoured cod but not capelin.

Strong and medium strength year classes of cod prior to 1970 would have been accompanied by good year classes of herring, providing plentiful food for the older piscivorous cod. This is emphasized by Izhevskii (1961): "everything that has been said about cod applies . . . to herring. This means that trends in the appearance of abundant year classes of herring, growth of populations and biomass, are identical with those of cod. This is accounted for by the fact that the biological and inorganic processes . . . which control the reproduction of the herring, are elements of the same system in which both cod and herring live, i.e. the Atlantic system". Historical data about the geographical distribution of young herring in the Barents Sea (Dragesund and Gjøsæter, 1988; Izhevskii, 1961, 1964; Konstantinov, 1967) demonstrate that a CCA ordination for herring would have placed it within the "cod-haddock" domain. As expected, the 1983 year class of herring was very strong as 0-group but the 1984 and 1985 year classes were weaker, partially due to predation from the large numbers of young cod in the Barents Sea in the mid-1980s (Anon., 1991). Lack of herring, and a spatially and temporally out-of-phase cod-capelin cycle increases cannibalism of cod.

There has been much discussion about the cause-and-effect relationship leading to the decline in several of the dominant fish stocks of the Barents Sea. Proposals regarding the decline of the capelin have encompassed over-fishing, predation by the 1983 year class of herring as well as inter-specific competition between these two species, increased predation by the strong 1983 year class of cod, and variability in climatic domains (Hamre, 1985; Skjoldal and Rey, 1989; Hopkins and Nilssen, 1991). There is little doubt that several of these factors have contributed towards the problems. The results of the present study, however, clearly emphasize the differing environmental responses of the four key species (cod, haddock, capelin, and polar cod) and of course herring ("the missing link") in the advective hydrographic system of the Barents Sea.

Future aspects of our work will incorporate food/

stomach contents analyses of cod (Mehl, 1989), as well as demographic information for the various species. This will offer increased possibilities for quantifying intra- and inter-specific aspects of vital demand–supply relationships (Ney, 1990).

Finally, we draw attention to a passage written over 40 years ago (Lee, 1949): "It may be possible to forecast the future trend of Arctic climate by obtaining correlations between variations of the meteorological and hydrographic elements in the Arctic and those in other parts of the world. ... If we seek, therefore, to forecast variations by means of periodicities or correlation, we must adopt sound statistical methods and severe standards. By studying the interaction between solar radiation and the atmospheric and oceanic circulations along lines such as these, oceanographers may be able to make a most significant contribution to the investigation of the mechanism of climatic variations to fisheries research." Such thoughts seem particularly apt today.

Acknowledgments

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Availability of blue whiting (*Micromesistius poutassou*) in Faroese waters in relation to hydrography

Bogi Hansen and Stein Hjalti í Jákupsstovu

Hansen, B., and Jákupsstovu, S. H. í. 1992. Availability of blue whiting (*Micromesistius poutassou*) in Faroese waters in relation to hydrography. – ICES mar. Sci. Symp., 195: 349–360.

Postspawning blue whiting migrating from the spawning areas west of the British Isles to the feeding areas in the Norwegian Sea are fished from late April to early June in Faroese waters, mainly along the southern parts of the Faroe Plateau. The main migratory route is through the Faroe–Shetland Channel. In some years, however, when a large proportion of blue whiting migrates through the Faroe Bank Channel, the success of the fishery is significantly better owing to the narrower shape of the Faroe Bank Channel compared with the Faroe–Shetland Channel. Data are presented for the last ten years on the fishery, including the distribution of blue whiting in the area from acoustic surveys and the distribution of the main watermasses from CTD profiles obtained during the acoustic surveys. It is found that the blue whiting distribution is consistent with relatively passive drifting and that the fish drift west of the Faroe Plateau through the Faroe Bank Channel when there is a strong flow towards the west in the Faroe–Shetland Channel at the preferred depth.

Bogi Hansen and Stein Hjalti í Jákupsstovu: Fiskirannsóknarstovan, Nóatún, 100 Tórshavn, Faroe Islands.

Introduction

The main spawning area of the northern stock of blue whiting is along the shelf edge west of the British Isles in March to April. After spawning, the blue whiting migrates north to the feeding areas in the Norwegian Sea (Fig. 1). The migration past the Faroes is partly through the Faroe–Shetland Channel (FSC) and partly through the Faroe Bank Channel (FBC), (Bailey, 1982). Since commencement of the large-scale blue whiting fishery in the late 1970s the main fishing in Faroese waters has been from late April to early June. The success of the fishery in Faroese waters in this period has to a very large extent been governed by the geographical distribution of the fishable concentrations. The fishery has been most successful in years when these are found in the FBC. The distribution of the fishery in Faroese waters has varied irregularly with no apparent relation to the size of the spawning stock, and it has been hypothesized that the changing horizontal distribution is related to changes in hydrographical features (Hansen *et al.*, 1979). The main aim of this paper is to discuss this question.

Materials and methods

In this paper information from different sources is combined. One of these sources is a series of combined echo-acoustic and hydrographic research vessel surveys. In addition, information on the fishery, partly from the Faroese Statistical Office and partly from the Faroese Coast Guard, is used.

Echo-abundance surveys

Up until 1981 the research vessel available to the Fisheries Laboratory of the Faroes was the "Jens Christian Svabo". This was a fairly small vessel unsuited to mid-water trawling for blue whiting. It was, however, equipped with a Simrad Ek 38 echo sounder connected to a Simrad integrator. Since 1981 the research vessel "Magnus Heinason" has been used by the Laboratory. This vessel is suitable for pelagic trawling to a depth of approximately 500 m. It is equipped with an Ek 400 Simrad echo sounder connected to an integrator.

Using these two vessels annual echo surveys on blue whiting have been carried out in April–May since 1977.

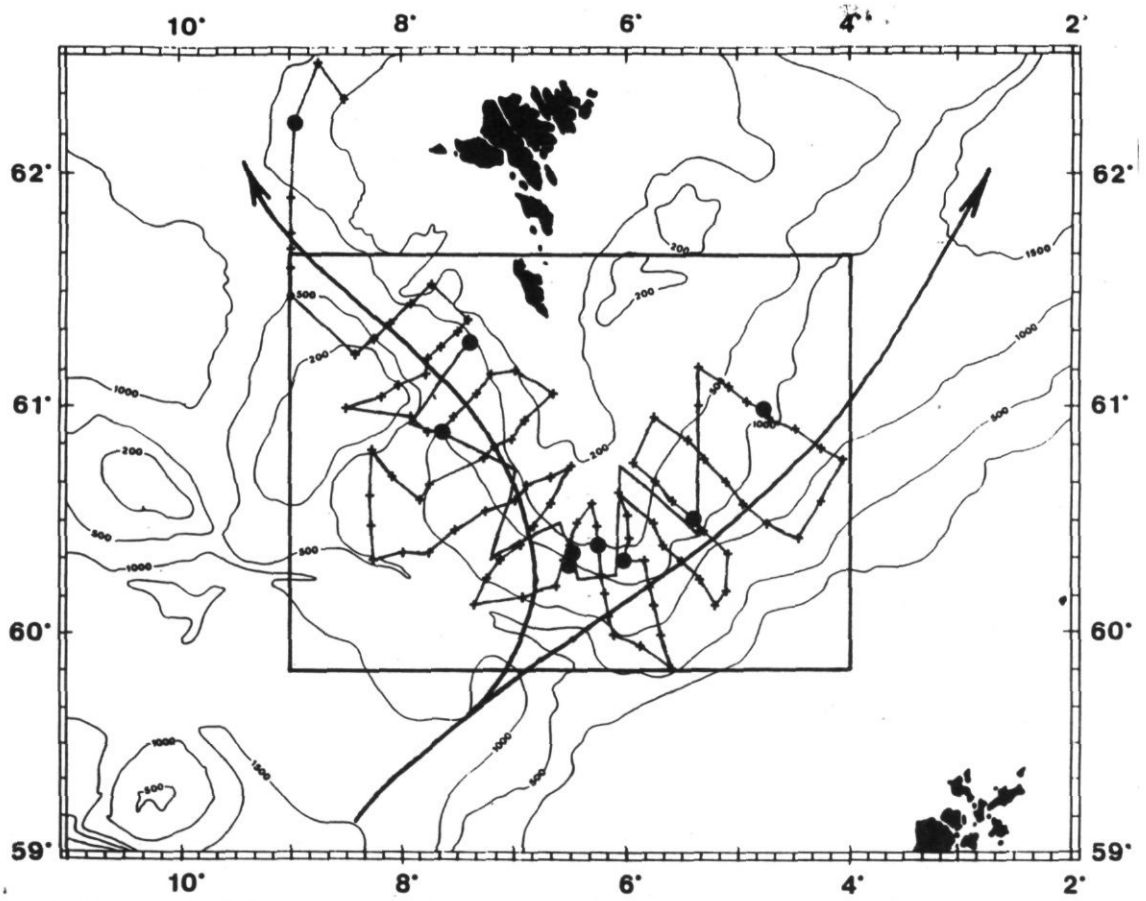


Figure 1. The area studied. Bottom topography is shown by isobaths. The cruise track of the 1988 combined echo acoustic and hydrographic survey is shown by the continuous track. The plus signs on the track denote hydrographic CTD stations and the filled circles are pelagic trawl hauls. The area in the inset rectangle is the one shown in Figures 2 and 6. The arrows indicate the two alternative migration routes of blue whiting through Faroese waters.

The purpose of the surveys has been to monitor the post-spawning migration of blue whiting past the Faroe Plateau. Owing to technical difficulties only the surveys from 1982, 1984, 1985, 1986, 1988, 1989, 1990, and partly 1987 were sufficiently complete to be analysed in the context of this paper. There has been some variation from one year to the other in the actual survey grid, but in recent years the surveyed area has been almost the same, as exemplified by the cruise track in Figure 1.

During the surveys the echo intensities have been allocated to species based on pelagic trawling. In the years when the "Jens Christian Svabo" was used this was done by using information from the fleet fishing for blue whiting in the area. Since the "Magnus Heinason" was taken into use, however, regular trawling has been part of the cruise schedule (Fig. 1).

Before the survey the echo sounders were calibrated using a standard copper sphere. As the scope of this paper is to show the relative distribution of the blue whiting migration, only echo values are given, not translated into biomass. The echo values reported are averaged over five-mile cruising intervals.

Hydrographic observations

On both the research vessels used for the echo surveys a Neil Brown CTD was available; up to 1989 a Mark III and in 1990 a Mark V. Hydrographic CTD observations were taken regularly during the echo surveys and thus simultaneous values for fish abundance and hydrographic features can be compared. During the earlier years the number of CTD stations was limited; it soon became apparent that the hydrography of the region was so complicated that a dense two-dimensional net of stations was necessary for mapping the temperature and watermass distribution adequately. The station net shown in Figure 1 is typical of the later years.

The CTD data were processed and averaged to 1-m depth intervals (Hansen, 1992). In later years, water bottle samples were taken on each CTD profile for

salinity calibration on an Autosal salinometer ashore. On the first surveys only a very limited number of calibration samples were obtained; but the long-term stability of the Deep Norwegian Sea Water allows frequent checks of the salinity values, and for the purpose of this work the accuracy is quite adequate.

Fishery information

All large Faroese fishing vessels are required by law to keep fishing log-books and although the reporting is not complete it should be sufficiently representative to give the geographical distribution of the fishery. The log-book data are kept in a database at the Faroese Statistical Office (Veiðihagtalsstovan). From this database all trawl hauls aimed for blue whiting by the Faroese blue whiting fleet in May have been sorted by year and by statistical rectangle, and the average catch per unit effort (c.p.u.e.) (tonnes/hour) by statistical (log-book) rectangle calculated.

When the fishery takes place on the western slope of the Faroe Plateau the fishable concentrations are confined to a narrow and often very dense band along the shelf edge on which the international fleet concentrates its fishery. In order to prevent chaos and, consequently, dangerous situations, the Faroese Coast Guard is legally allowed to organize the fishery in specific areas. In practice they define an area within which the trawlers are allowed to trawl in specified directions, usually parallel to the shelf. In years when the Coast Guard finds it necessary to regulate the fishery in the FBC it is reasonable to assume that a significant proportion of the post-spawning blue whiting migrates through this channel.

Results

Echo surveys

In Figure 2 the main results from the echo surveys are presented for the eight years with sufficiently good data. Shown for each year are the average integrated echo densities allocated to blue whiting and temperature at 400-m depth. The echo-density values have also been summed over depth and averaged in statistical rectangles (1° longitude × 1/2° latitude). The rectangles have then been grouped into a western subarea and an eastern subarea, as shown in Figure 3. Summing the echo intensities of the rectangles an echo intensity value was derived for each subarea and, in accordance with whichever of the two values was the larger, the distribution for each year was characterized as either: W (western) or E (eastern). In 1987 the echo integrator did not work during the survey of the eastern part of the FSC. From qualitative grading of the echo traces, however, only insignificant concentrations were found in this area compared to the FBC. The results are presented in Table 1.

Log-book reports

Two examples of results from log-book reports are shown in Figure 3. The Figure shows total catch reported in each rectangle for two years (1982 and 1988) which have clearly different geographical distribution of the catch. The thick line divides the area into the same subareas (W and E) which were used for the survey data.

The numbers in parentheses below the catch numbers are the c.p.u.e. in tonnes per trawlhour. These are shown only for rectangles with catches above 1000 tonnes, but they illustrate very clearly the difference in catchability between eastern and western years. Although the total catch reported was slightly higher in 1982 than in 1988, the c.p.u.e. was generally much higher in 1988, when the distribution was western, than in 1982, when it was eastern.

These results have been obtained for each year since 1977 except for 1983, when they were not deemed reliable. For each year the total catch was evaluated for each of the two subareas and, based on this, the year was categorized as either western (W) or eastern (E). The result is shown in Table 2.

Coast Guard regulations

Table 3 gives the years when the Faroese Coast Guard put into effect regulations governing the blue whiting fishery in the FBC. As mentioned previously, regulation indicates a western distribution of the fishery (information is missing for 1984).

Discussion

The horizontal distribution

Tables 1, 2, and 3 present three different evaluations of the annual distribution of blue whiting. When comparing these tables the three independent methods estimating the distribution give similar answers to an amazing degree. Especially noteworthy is the corre-

Table 1. The classification of years into eastern (E) or western (W) distribution of blue whiting based on echo survey data.

	1982	1984	1985	1986	1987	1988	1989	1990
Distribution	E	E	W	E	W	W	W	W

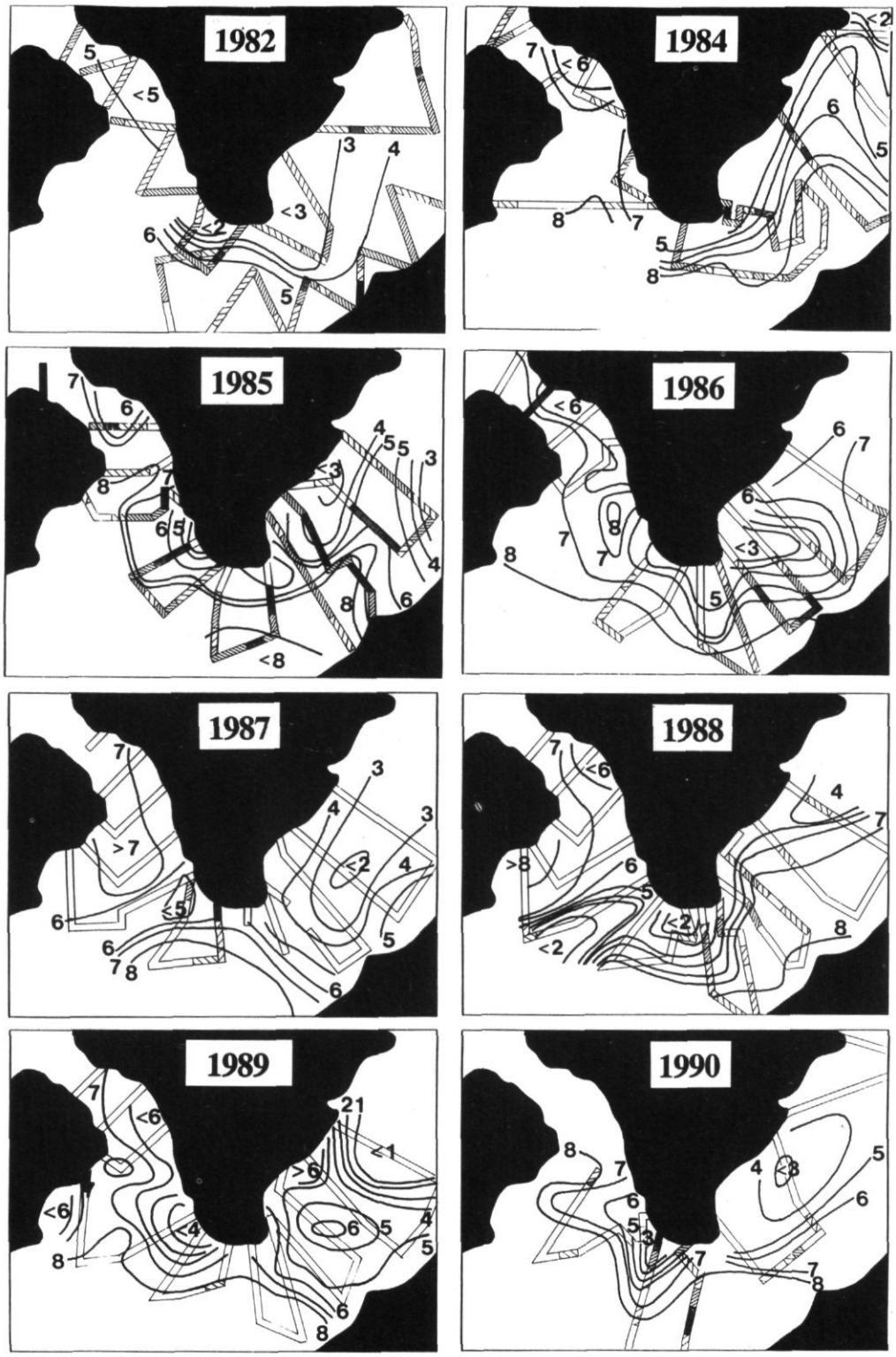


Figure 2. Temperature at 400-m depth and blue whiting abundance during the annual echo surveys from 1982 to 1990 (except for 1983). The acoustic integration track is shown by the parallel lines with relative echo intensity indicated by the hatching between them in the scale shown below. The continuous lines show temperature at 400-m depth (in °C).

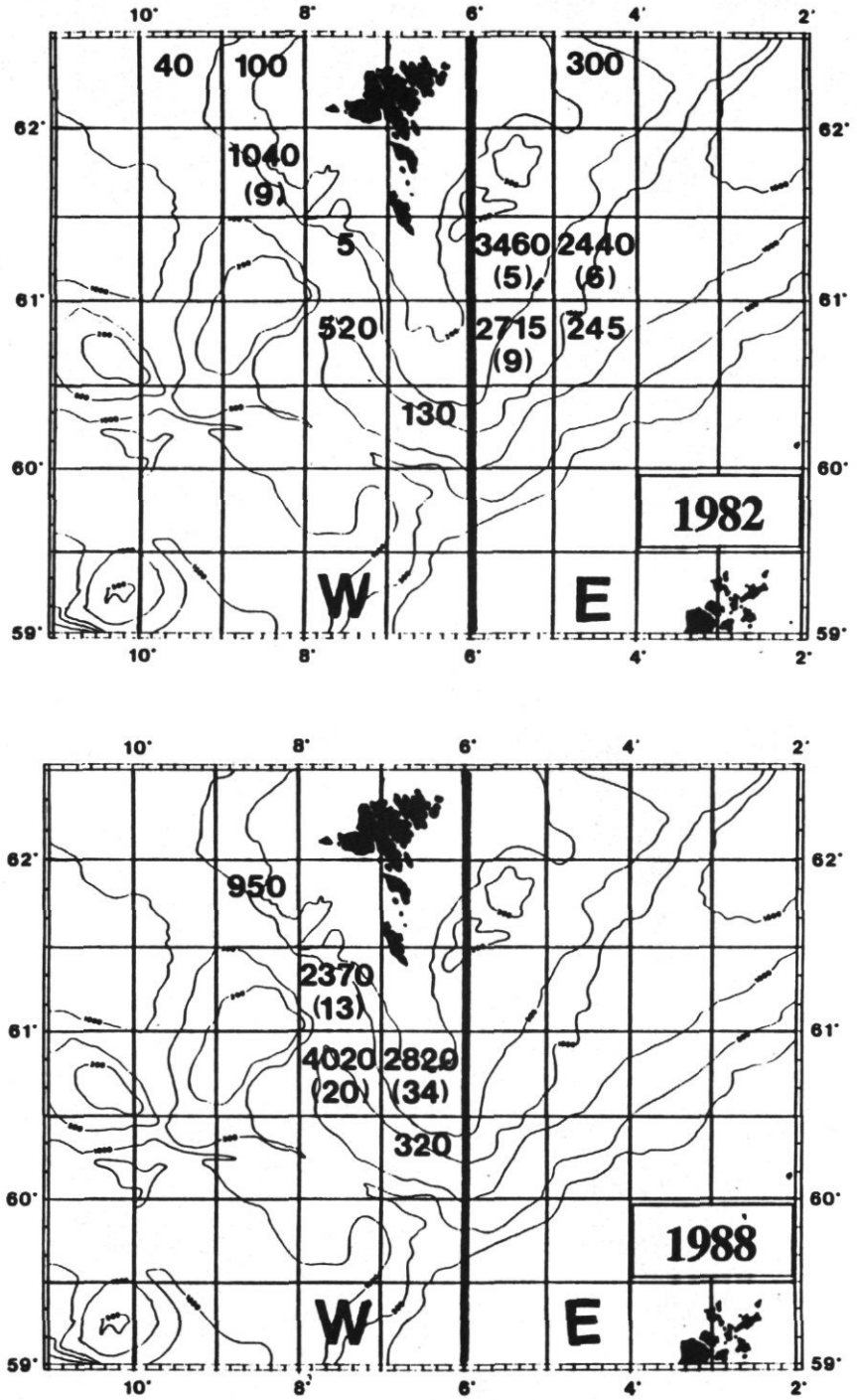


Figure 3. The reported catch of blue whiting in tonnes in statistical rectangles in May in two different years, one with a typical eastern distribution of the catch (1982) and one with a typical western distribution. Numbers in parentheses indicate catch per unit effort (tonnes/hour) for the rectangles with catch above 1000 tonnes. The thick vertical line divides the region into a western (W) and an eastern (E) subarea.

Table 2. The classification of years into eastern (E) or western (W) distribution of blue whiting based on reported catch.

	1977	1978	1979	1980	1981	1982	1984	1985	1986	1987	1988	1989	1990
Distribution	W	E	W	W	W	E	E	W	W	W	W	W	W

Table 3. Years since 1980 when the Faroese Coast Guard did (+) or did not (−) regulate blue whiting fishery west of the Faroes.

	1980	1981	1982	1983	1985	1986	1987	1988	1989	1990
Regulation	+	+	−	−	+	−	−	+	+	+

spondence between the survey distributions, which show abundance, and the log-book distributions, which show catch. As a western abundance distribution is more fishable than an eastern one, one could expect some years with eastern survey distribution, but western log-book distribution. Thus the discrepancy in 1986 is not unexpected.

The similarity gives credence to the results obtained for the distribution by each separate method, but it also leads to further conclusions. One is that the abundance distribution obtained from the echo surveys generally represents the distribution of the fishery, too, which usually extends over a longer period. The corollary is that *if the abundance distribution from the echo surveys can be related to hydrographic features then these features also have a profound influence on the fishery*. The rest of this discussion therefore concentrates on relating the abundance distribution to the hydrography.

The vertical distribution

During the echo surveys the echo intensities are summed and averaged in integrator channels representing depth intervals. Usually 50-m intervals have been used at those depths where the blue whiting is mostly found. Blue whiting has a distinct diurnal vertical migration cycle, but during daytime the densest concentrations are generally found well below the surface. The depth of maximum concentration varies, but in this region 400 m is a typical value.

A pertinent question is whether this preferred depth is determined by temperature or by some other hydrographic parameter. In Figure 4 some typical sections showing fish abundance and temperature do not seem to indicate any large influences from temperature on the vertical distribution. Blue whiting apparently do not like water colder than 2°C, as has been previously noted in other areas (Blindheim and Jákupsstovu, 1976); but in this material we see no relationship between preferred depth and temperature, and the large slopes of isotherms in some of the sections are not reflected in depth changes of the fish concentrations.

The horizontal distribution of temperature and watermasses

The above results were the rationale for plotting fish abundance together with temperature at 400-m depth in Figure 2. Although the fish are found at other levels, the temperature at this depth should be fairly representative of the environment of the blue whiting during its migration through this area.

The most prominent feature in the temperature distributions in Figure 2 is the front usually found south of the southern tip of the Faroe Plateau. This front is not very distinguished at the surface where the temperature differences are small, but at mid-depths it is usually more pronounced. The front, termed here the midwater front, separates the waters of western and eastern origin. The waters from the west mainly consist of Modified North Atlantic Water (MNAW) down to some 500 m (Becker and Hansen, 1988), where it is underlain by Deep Norwegian Sea Water (DNSW) flowing out through the FBC. In the surface the waters coming from the east also consist mainly of MNAW, having passed the Iceland–Faroe Ridge north of the Faroes; but in the Iceland–Faroe Front north of the Faroes water of more arctic origin is sinking. This water is termed NI/AI water (North Icelandic/Arctic Intermediate) and is mixed into the MNAW so that the eastern water in the FSC decreases much faster in temperature with depth than the western water in the FBC. This is illustrated in Figure 5.

A second noteworthy feature in Figure 2 is the temperature difference across the FBC. In all the years shown in Figure 2 the distribution at 400 m shows the occurrence of colder water along the eastern flank of the FBC. In 1986 a continuous band was observed, although this to some extent could have derived from inadequate sampling resolution. Usually, however, the colder water over the Faroe slope occurs in boluses with typical dimensions of some 30 miles.

There are two possible explanations for this colder water along the eastern flank of the FBC. At greater depth the channel is filled by DNSW with temperatures below zero, and the cold water along the eastern flank of

the FBC could be due solely to a sloping of the isotherms across the channel. This may well be part of the explanation, but in addition there is evidence of a larger component of NI/AI water along the eastern flank of the FBC than on the other side of the channel. This can be seen from the salinity. In addition to being colder the NI/AI water is considerably less saline than the MNAW and also the underlying DNSW. In a TS diagram the influence of the NI/AI water is thus seen as a curvature with reduced salinity at the source temperatures (2–4°C) of the NI/AI water. This can be seen in Figure 5, which shows one typical western station in the FBC and a station in the FSC with strong influence of NI/AI water. The figure also shows a station in one of the boluses and the intermediate character of the TS trace for that station indicates fairly large amounts of NI/AI water. The example shown in Figure 5 is not normal, as the NI/AI water content in the FBC is usually low, but still it can be seen from Figure 6 that at a temperature of 4°C the cold water boluses generally have salinities 0.02 to 0.06 lower than the rest of the FBC. The NI/AI water in the FBC thus has been well diluted compared to the FSC, where salinities below 35.00 are found at 4°C, which again is well diluted compared to the source salinities north of the Iceland–Faroe Front which are usually below 34.8.

In spite of this dilution, the existence of NI/AI water in the cold boluses in the FBC is an indication that they originate in the region of the midwater front and are being advected towards the north along the slope of the Faroe Plateau. At present the observational material does not allow a complete understanding of the circulation of the FBC, but for the northern part of the FBC between Faroe Bank and the Faroe Plateau the influence of NI/AI water is much less evident along the western flank compared to the eastern. This could indicate that the water in the western part of the channel derives from the north. This is consistent with the very persistent anticyclonic circulation over the shallower parts of Faroe Bank (Hansen *et al.*, 1986) and with the movement of satellite tracked drogues in the channel (Hansen *et al.*, 1991).

The flow field

A tentative picture of the general flow field at the preferred depth of the blue whiting seems to show three branches of water meeting in this region. The *Southern branch* comes from the Rockall Trough over the Wyville Thomson Ridge. Flowing southward along the western flank of the FBC is the *Northern branch*. These two branches mostly consist of MNAW and stay on the western side of the midwater front. The *Eastern branch* flows southwestward along the eastern slope of the Faroe Plateau. It has NI/AI water admixed into its MNAW origin and stays mainly east of the midwater front, but this branch releases water through the front in the form of eddies or boluses which are mixed with water from the Northern and possibly also the Southern branch and flow northwards along the eastern flank of the FBC.

From Figure 2, however, it is evident that there are significant differences in the flow pattern from one year to the next. Of special interest is the east–west component of the flow at the depth of the blue whiting. The question arises whether one can group different years according to the flow. Is it possible to distinguish between eastward (E) years when the flow at mid-depths has a comparatively strong east-going component and westward (W) years when the flow is less strong towards the east or even westward? Lacking direct current measurements, watermass analysis might indicate variations in this flow component since the NI/AI water in the FSC is transported into the channel by this flow and large amounts of this water should indicate comparatively strong westward flow. The source characteristics of NI/AI water are, however, in themselves fairly variable and at the present stage a complete watermass analysis is not possible.

But, as argued above, the salinity in the FSC at 4°C can be considered as an indicator of NI/AI water by itself and therefore also as an indicator of westward flow at mid-depths. Taking the location of the 35.00 isohaline for 4°C water in Figure 6 as the criterion, in 1985, 1987, 1989, and 1990 the 35.00 isohaline at 4°C extended far west into the FSC, indicating strong westward flow at mid-depths compared to the years 1982, 1984, and 1986, when the 35.00 isohaline at 4°C was found only in the far northeastern part of the FSC, if at all. The salinity at 4°C thus does provide a method for distinguishing between E and W flows during the survey periods for these years. In 1988 some low salinity water was found close to the Scottish slope but it did not extend across the channel and it is difficult to classify that year by this method.

Details about the mechanisms underlying the variations of the east–west component of the flow at mid-depths are not known, but presumably these variations reflect variations in the balance between the three current branches meeting in this region.

Blue whiting distribution and hydrography

It appears from Figure 2 that the blue whiting is mostly found on the western or southern (the Atlantic) side of the midwater front, with the largest concentrations in the frontal region. With a few exceptions only small concentrations are found well east of the front in the Eastern branch. It may be noted that the same behaviour has been observed in the feeding area in the Norwegian Sea where the blue whiting also seems to concentrate close to the Iceland–Faroe Front, but on the Atlantic side of it. The same behaviour is indicated in Figure 4, where the largest concentrations are found in the frontal regions. This is most clearly seen on section

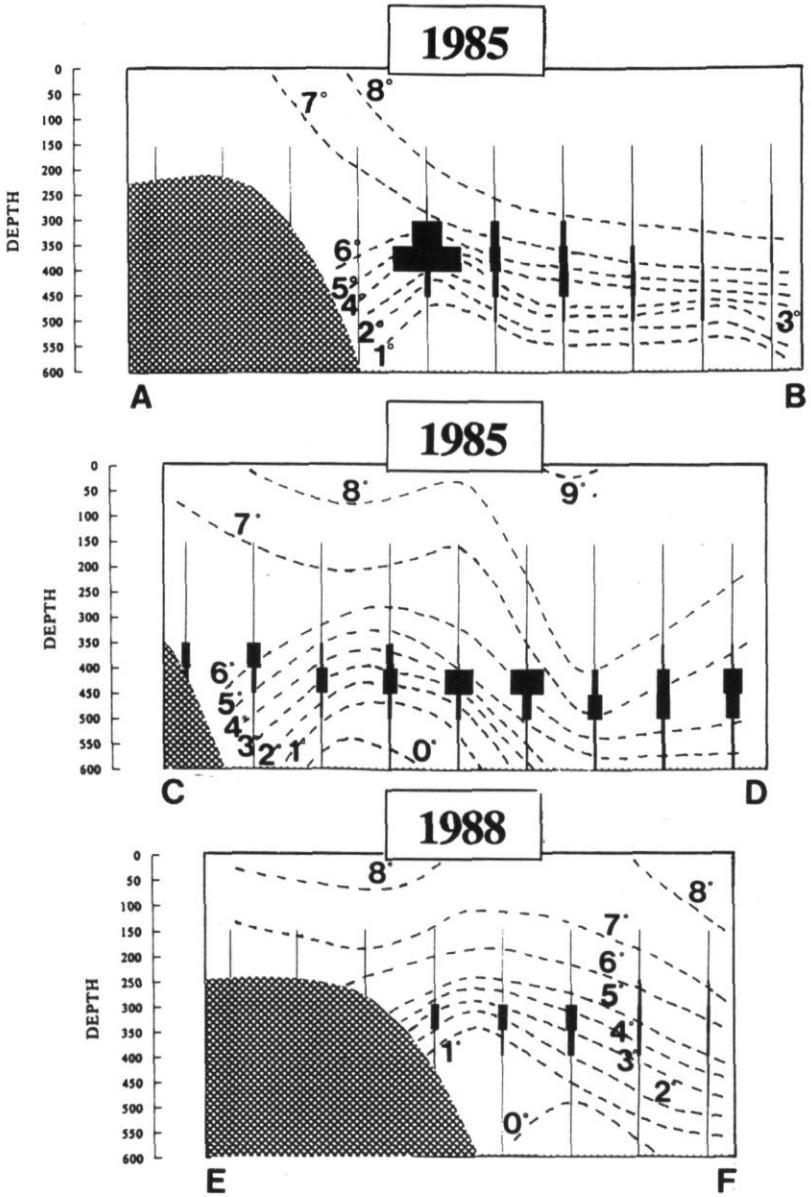
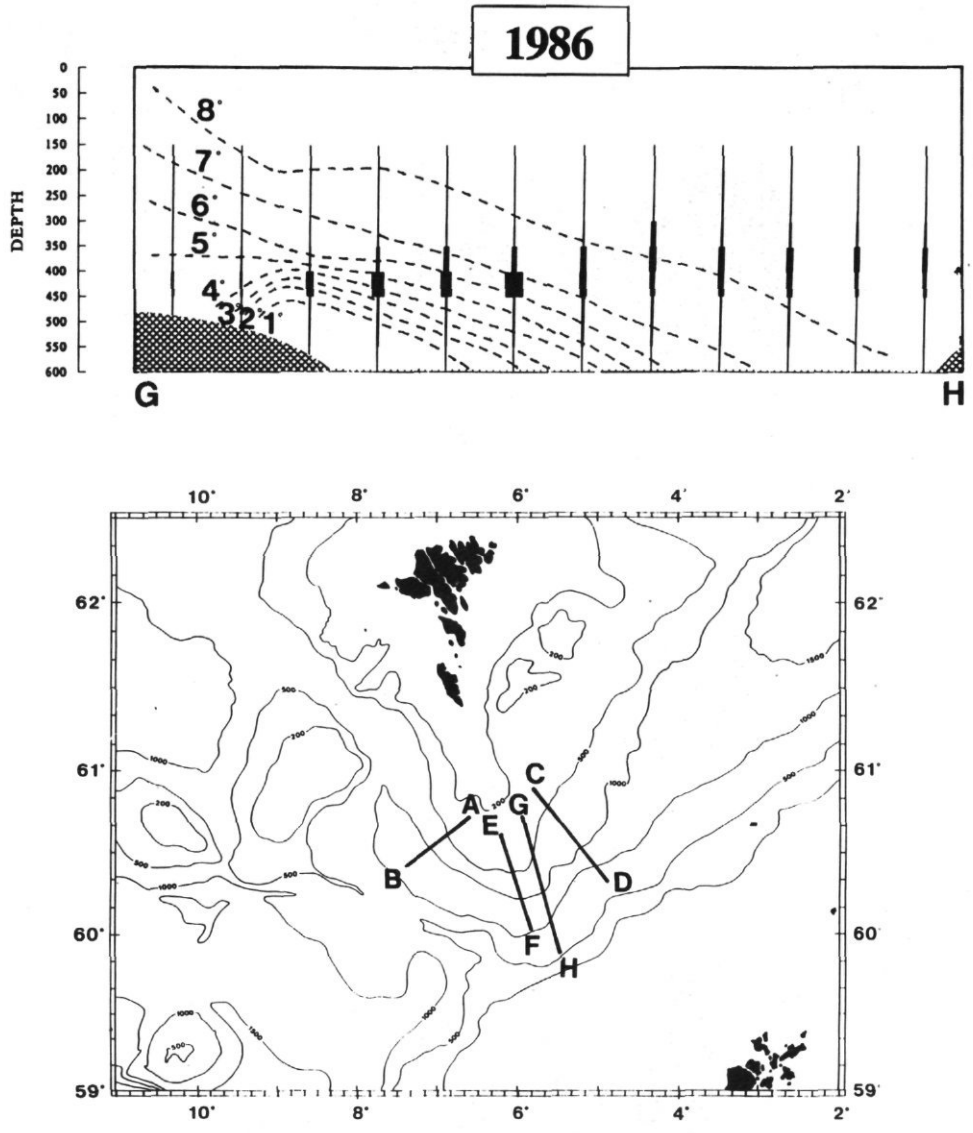


Figure 4. Typical vertical sections of blue whiting abundance and temperature in different years. Temperature is shown by the dashed isolines (in °C). Fish abundance is indicated by bars, whose width is proportional to echo intensity. The map shows section locations.

AB 1985. This distribution is consistent with a more or less passive advection of the blue whiting, which is reasonable to expect after spawning when the blue whiting is weak. In addition, there may be some dynamical or biological mechanism concentrating the fish in the front.

The timing of the annual echo surveys is about one-and-a-half months after the beginning of the spawning period, and the distance from the spawning areas to the southern tip of the Faroe Plateau is in the order of 500 km. Passive advection thus requires current speeds in the order of 10–20 cm/sec, which does not seem unrealistic especially if part of the transport is in the Continental Slope Current. The hypothesis does not, of course, exclude active swimming, only the fish seem to follow the flow directions.

If this hypothesis holds true, the years with comparatively strong eastward flow at mid-depths would be expected to give a more eastern distribution of the fish than the years with less strong eastward or even westward flow. In the previous paragraph we used the hydrographic observations to group the years into E or



W years for the flow and in Table 4 we compare this grouping with the fish distribution as determined by the three different methods (Table 1, Table 2, and Table 3). The year 1988 was assigned a question mark as it was not possible to classify it by the chosen scheme.

The correspondence between the fish distribution and hydrography in Table 4 is amazingly good, especially if the fish distribution determined from the acoustic surveys is used, which should also be the most appropriate choice. It is therefore probable that it is the east–west

Table 4. The classification of years into eastern (E) or western (W) blue whiting distribution based on three different sources of information and the classification of the flow at mid-depths into eastward (E) or westward (W) based on hydrographic observations.

	1982	1984	1985	1986	1987	1988	1989	1990
Distribution from survey	E	E	W	E	W	W	W	W
Distribution from logb.:	E	E	W	W	W	W	W	W
Distribution from regulation			W			W	W	W
Flow direction	E	E	W	E	W	?	W	W

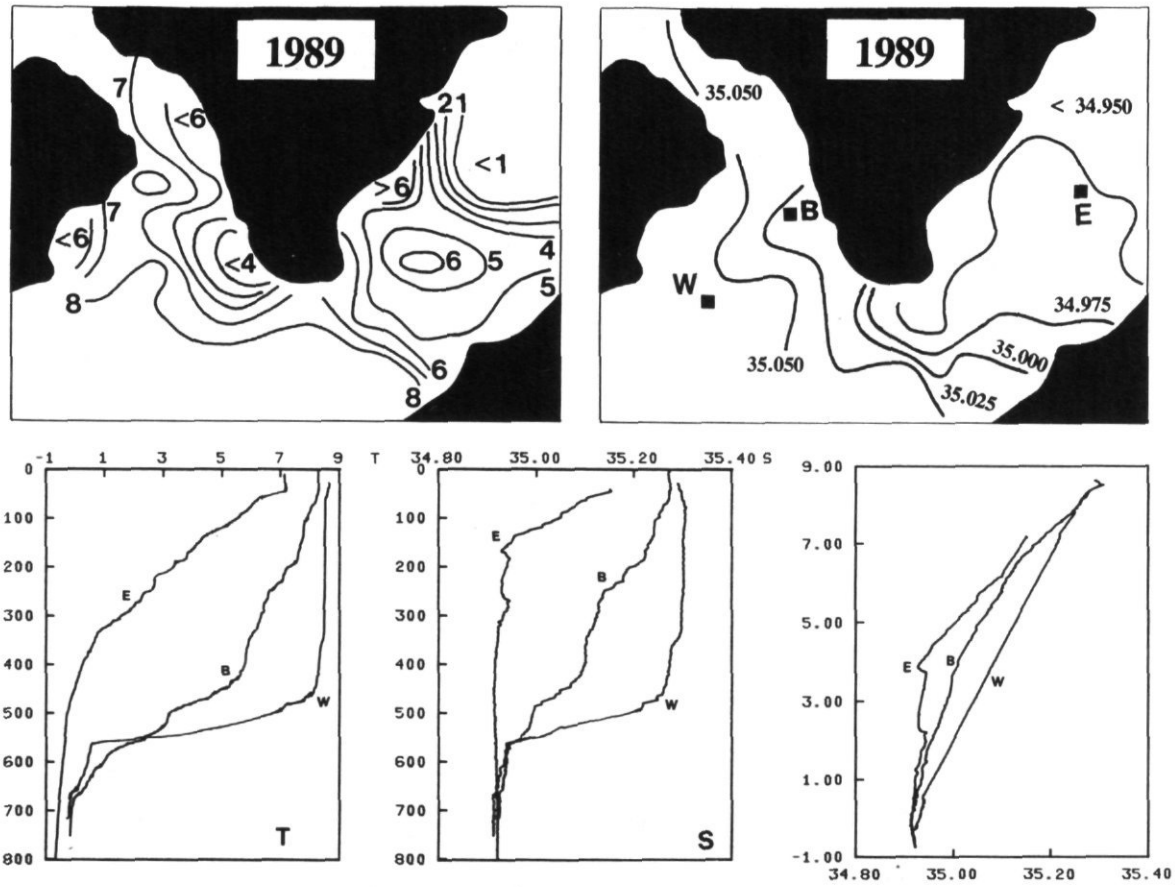


Figure 5. The different types of water in the upper parts of the Faroe channels. The maps show temperature at 400-m depth (upper left) and salinity of water with temperature 4°C (upper right) in 1989. The salinity map shows three locations: one in the western (W) water (MNAW), one in the eastern (E) water and one in a bolus (B) containing a mixture of both. Profiles of temperature (lower left) and salinity (lower middle) and TS diagrams (lower right) are shown for each of these three locations.

component of the flow at mid-depths in the FSC which to a large degree determines the migration path of the post-spawning blue whiting past the Faroes.

At present this is a working hypothesis. If further research supports it the major efforts now being directed into observing and modelling the major ocean currents might in the future permit prediction of the blue whiting migration in this region.

A further perspective is the question concerning how much of the blue whiting at the spawning fields drifts towards the Faroes and the Norwegian Sea. With a more or less passive drift a weakening of the northeastward flow of Atlantic water through the Rockall Trough the transport of fish towards the northeast should also decrease. In 1987 Figure 2 indicates blocking of the Southern branch both east into the FSC and north into the FBC, thus only small numbers advected into the Faroe area. Indeed the echo survey showed exceptionally small values that year (Fig. 2).

This relates to the question recently raised (Monstad, 1990) that significant proportions of the northern blue whiting spawning stock in later years apparently have not migrated into the Norwegian Sea, and thus might have been advected into more oceanic areas rather than through the Rockall Trough.

Conclusion

Summarizing the results obtained, the echo surveys seem to give a good representation not only of the instantaneous blue whiting biomass distribution in the area, but also of its availability to fisheries (as documented by the consistency in Table 4). Furthermore, there is a clear connection between blue whiting concentrations and the watermass distribution, with the fish staying on the Atlantic side of the midwater front formed south of the Faroe Plateau by the confluence of waters from the west and from the east. This is consistent with a passive advection of blue whiting possibly com-

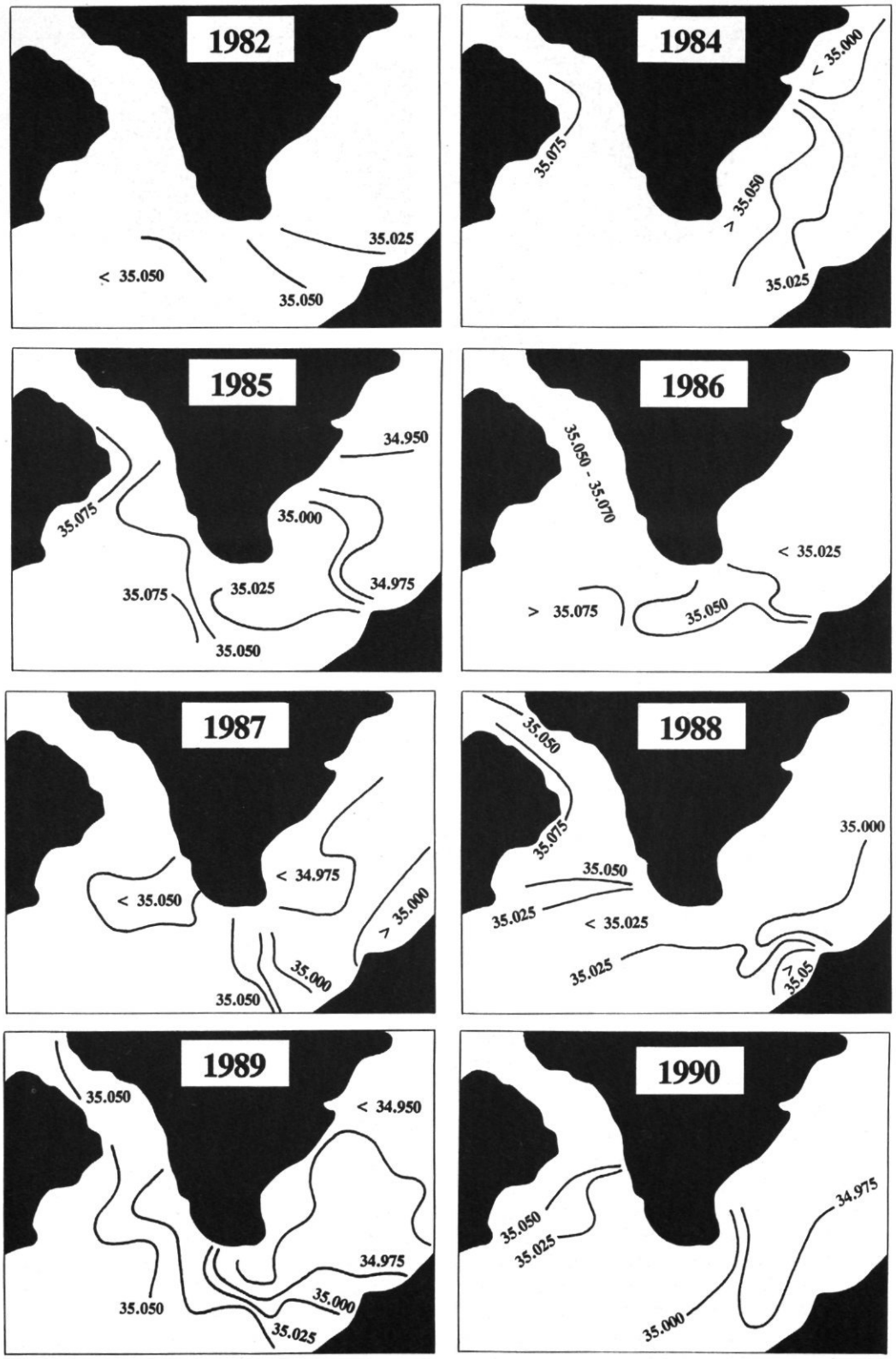


Figure 6. The salinity for water of temperature 4°C in different years.

bined with some mechanism for concentration in the frontal regions. In years when the hydrographic observations indicate a comparatively strong westward component of the flow at mid-depths this should give a more western distribution of the fish and Table 4 shows that this is indeed the case.

There are still many unanswered questions, however. The analysis of the data presented here is not yet complete and the observational effort is being continued in the hope that a more complete understanding of both the distribution and the total abundance in Faroese waters may be reached.

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Year-to-year variability in the plankton community on the spawning ground of the Hebrides–Buchan herring

Walter Nellen and Jürgen Schadt

Nellen, W., and Schadt, J. 1992. Year-to-year variability in the plankton community on the spawning ground of the Hebrides–Buchan herring. ICES mar. Sci. Symp., 195: 361–387.

Since 1967 the size of the herring stock spawning north of Scotland between the Hebrides and the Orkneys–Shetland Islands passage has been estimated in annual larval surveys by Denmark, Germany, the Netherlands, and the United Kingdom (Scotland). Plankton is collected by standardized Gulf III hauls and sorted for herring larvae. Samples taken by R.V. "Anton Dohrn" in September of 1982, 1983, 1984, and 1985 were analysed in greater detail for taxonomical composition, abundance of taxa, amount of biomass, and variability of parameters. The general composition of taxa did not differ much from year to year, but the abundance, total number of organisms, and amount of biomass expressed as dry weight varied. Cluster analyses, the geographical distribution of certain clusters, and the abundance of assumed indicator species showed that the influence of the North Atlantic drift on the spawning area varied and appeared to be greater in 1982 and 1983. In 1985 both zooplankton concentration and homogeneity were exceptionally high. Whether or not recruitment to the herring stock can be judged from such plankton analyses and, further, the extent to which they can help to clarify specific ecological situations are unsettled questions.

W. Nellen: Institut für Hydrobiologie und Fischereiwissenschaft der Universität Hamburg, Olbersweg 24, D-2000 Hamburg 50, Germany. J. Schadt: Fachberatung für das Fischereiwesen, D-8580 Bayreuth, Germany.

Introduction

The sea area to the north of the British Isles between northern Scotland, the Hebrides, the Orkneys, and the Shetland Islands (ICES Divisions IVa and VIa N) is an important spawning ground of herring, probably of the so-called Buchan stock. Each September, Denmark, Germany, the Netherlands, and the United Kingdom (Scotland) sample a grid pattern of stations (Fig. 1) for the newly hatched herring larvae. These surveys were proposed by ICES and have been carried out since 1967 to measure the size of the spawning stock (Schöfer, 1983; Saville and Rankine, 1984, 1985; Rankine, 1986). A standard plankton sampler, the Gulf III (Gehringer, 1952) or a modified version is used, allowing relatively precise determination of biomass per unit of sea surface area. Only herring larvae are sorted from the samples, though an additional routine analysis of the accompanying zooplankton would in the long term increase our knowledge of environmental variability and, perhaps, ecological changes in the epipelagic and herring recruitment processes, too.

Plankton samples collected within this program by R.V. "Anton Dohrn" (28 Aug to 12 Sep 1982, 8 to 18 Sep 1983, 6 to 16 Sep 1984, and 29 to 9 Sep 1985) have been closely examined to gain some insight into the

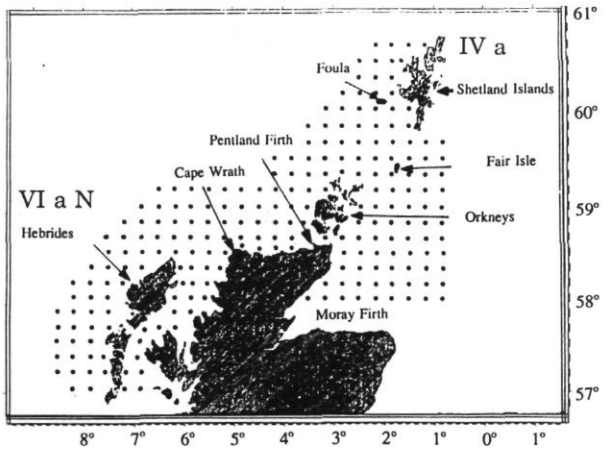


Figure 1. ICES station grid for the herring larvae census in September.

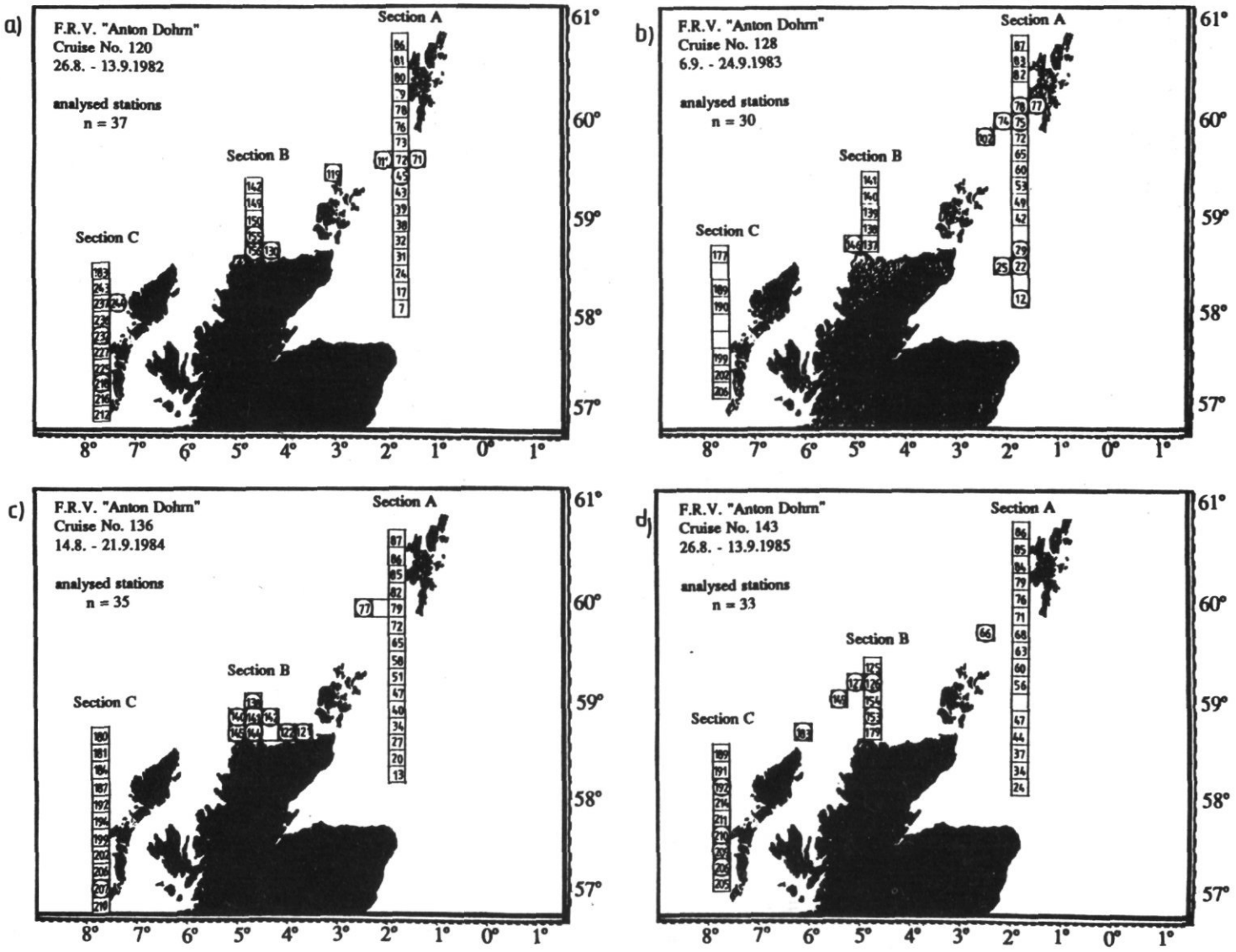


Figure 2. Positions and number of stations selected from the total zooplankton analyses.

structure of zooplankton composition and abundance and into the variability in time and space (Schadt, 1986). The sampling gear used was a "Nackthai" (Nellen and Hempel, 1969) with a mesh size of 300 μm.

Because lack of time and sorting capacity has not allowed processing and analysis of all of the 230 samples collected each year, three groups of stations were selected in an effort to reduce sample numbers (Fig. 2). In addition, the first ten stations with the highest concentration of herring larvae were analysed, even when they were not in one of the three sections shown in Figure 2. In total, the analysed samples amounted to 13% at minimum and 16% at maximum of all samples taken.

In order to count the plankton within a reasonable time, samples had to be split to between 1/10 and 1/25 of their original size. The degree of accuracy of counts was found to be ± 20% when at least 100 specimens were counted from each sample. Thirty-eight taxa were classified.

Similar plankton groups were classified by a hierarchical cluster analysis based on the Canberra-Metrik (Lance and Williams, 1966, 1967; Bölker *et al.*, 1980). The "representative plankton community" (RPC) for each defined cluster was determined by calculating the arithmetic mean of abundance per m^2 for each taxon. Yearly fluctuations of some taxa in section A were determined by calculating the standardized deviation (Williamson, 1961; Colebrook, 1963). Finally, the total biomass of zooplankton as dry weight was determined for each analysed sample.

Results

Distribution of herring larvae abundances

The area with high larvae abundance (more than 100 specimens per m^{-2}) is illustrated in Figures 3–6. High larvae concentrations were found west of the Hebrides, along the Scottish north coast from Cape Wrath to the Pentland Firth, and from the SW coast of the Shetland Islands to the Orkneys. Only in 1982 was larvae abundance high in all three areas. In 1983, larvae were absent from the Hebrides, but were abundant far south into the Moray Firth. In 1984 and 1985 herring larvae were not in higher abundance off the Shetlands. The eastern and western boundaries of larvae distribution were fixed at 1°10′ and 8°10′ W, respectively. Herring larvae abundance decreased considerably with larvae size.

Zooplankton

Presence of taxa

The abundance of some zooplankton taxa varied considerably from one year to the other (Table 1). Siphonophora occurred less frequently in 1983 and 1985, as did ctenophora in 1982, *Salpa fusiformis* in 1982 and 1985, *Doliolum* in 1984 and 1985, cladoceras in 1982, mysids in 1982 and 1985, *Eucalanus* in 1984 and 1985, and *Clione* in 1984. But most taxa were present regularly and occurred everywhere each year.

Abundance of taxa

To determine zooplankton abundance, single taxa were grouped in accordance with the principle that taxa of ecological similarity could be treated together (Table 2).

The abundance of total zooplankton and the abundance within groups is highly variable at all stations in all sections (Figs. 7–9). In sections A and B zooplankton density was generally low in 1982 and 1984, with an average of only about 40 000 specimens per m^{-2}. In the other two years average abundance was more than twice as high. Section C was poor in zooplankton in 1982 and 1983 compared to 1984 and 1985. Zooplankton concentration in the west was found generally to be low. This corresponds with the distribution of zooplankton biomass expressed as dry weight (Fig. 10); 1985 was especially rich in zooplankton in all three sections.

As Figures 7–9 show, copepods larger than 1000 μm and smaller than 1000 μm were obviously the dominating organisms. The third fairly abundant group comprised pteropods. While both size groups of copepods were numerous at all stations each year, the abundance of pteropods varied significantly, and just as much between stations as between years.

Only in 1985 were pteropods numerous and widespread at all stations in section A. In 1982 and 1983, only some of the stations had a high pteropod abundance, whereas in 1984 they were almost lacking. In general, this was the situation in section C, but the pattern in 1983 and 1984 was the reverse. Pteropods did not vary much in section B. For less abundant groups it is interesting that crustaceans other than copepods, i.e. mainly cladoceras, euphausids, and hyperiids, were always present in relatively high numbers. The more varying zooplankton forms were medusa, chaetognatha, tunicata, and herring larvae.

The first ten stations richest in herring larvae are arranged in descending order of larvae per m^2 in Figure 11. Zooplankton and herring larvae abundance are not correlated, but pteropods were present in all stations. Herring larvae and pteropod abundance, however, did not necessarily fluctuate absolutely in parallel (Fig. 12).

Cluster analysis

Clusters or groups were assumed to be different at the 50% level of resemblance (Fig. 13). This cut-off level or measure of similarity was not followed in 1984, when station nos. 79 and 82 were regarded not as a separate group but rather were included in one of the clusters nearest to these two stations. In 1985 distinct cluster

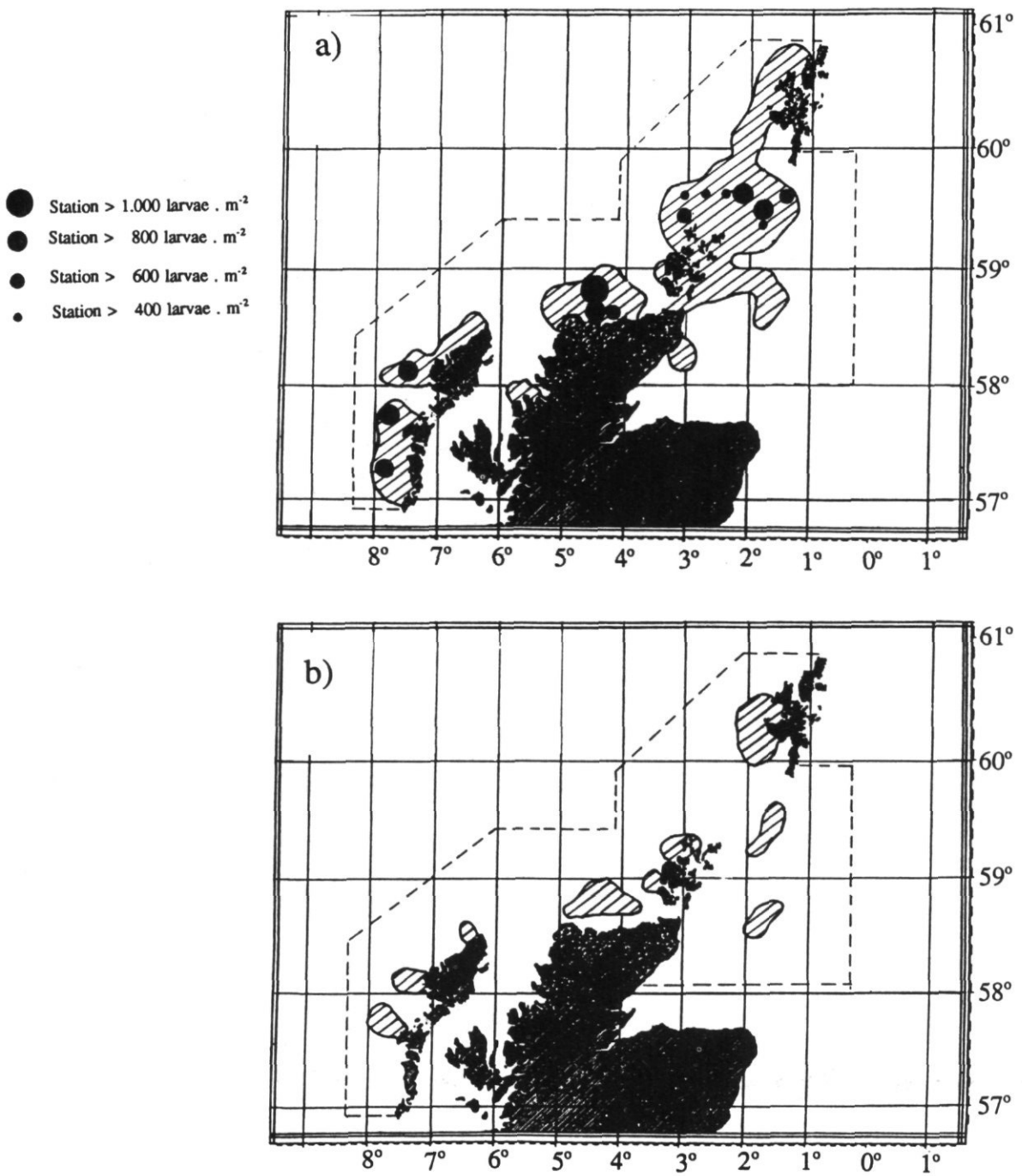


Figure 3. September 1982: (a) areas where herring larvae abundance exceeded $100 \cdot m^{-2}$; (b) areas where abundance of big herring larvae (>10 mm total length) exceeded $100 \cdot m^{-2}$.

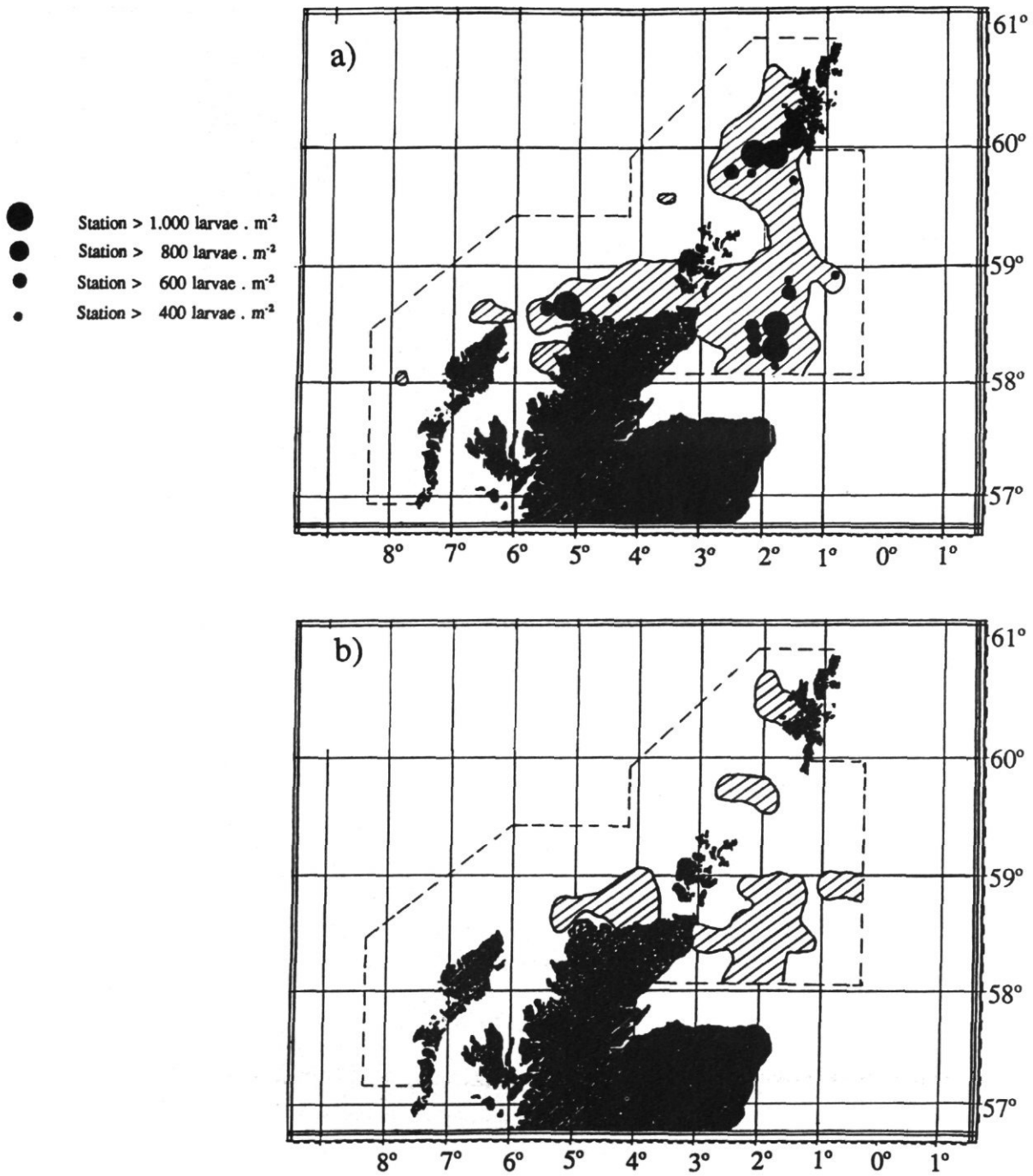


Figure 4. September 1983: (a) areas where herring larvae abundance exceeded $100 \cdot m^{-2}$; (b) areas where abundance of big herring larvae (>10 mm total length) exceeded $100 \cdot m^{-2}$.

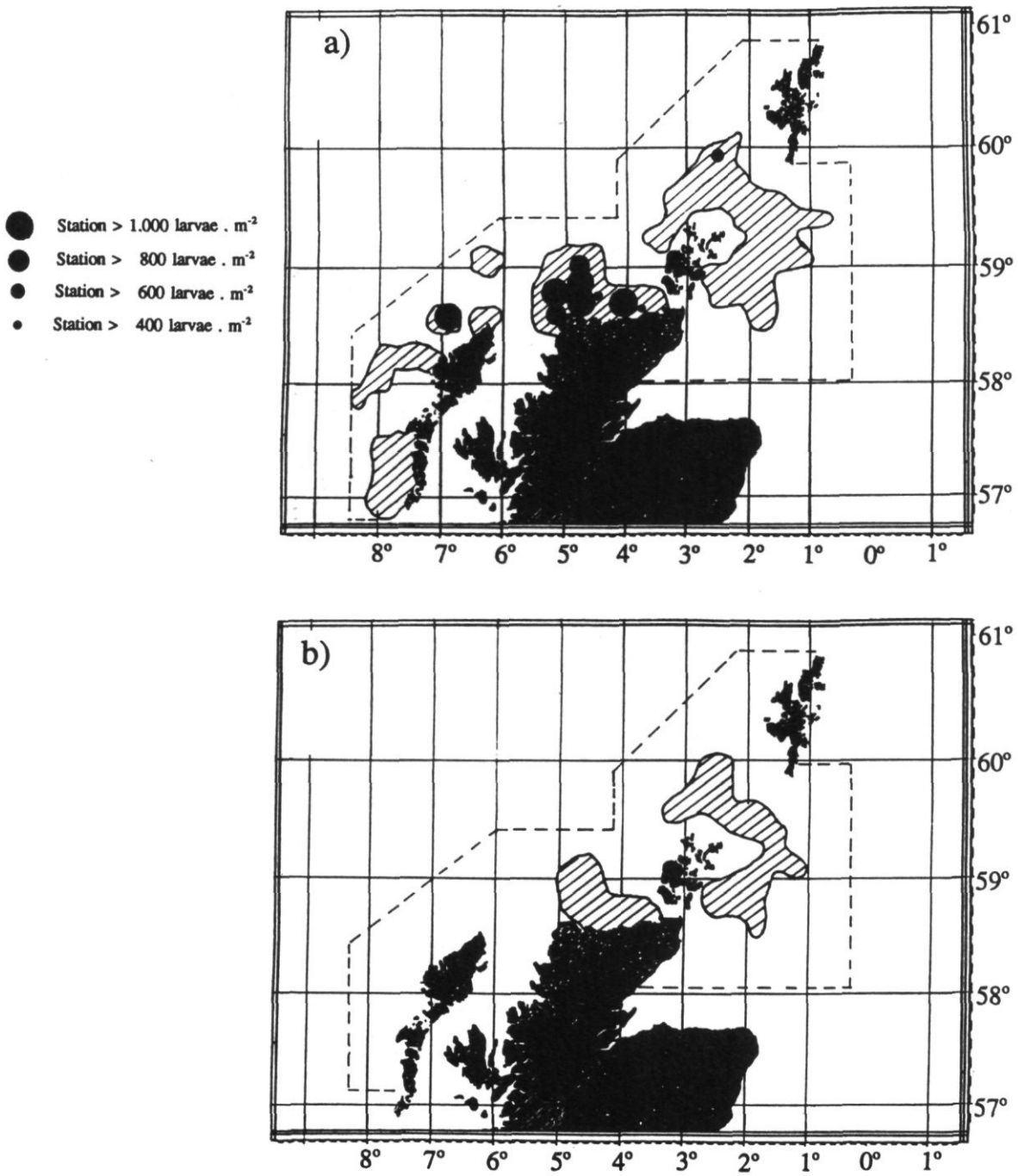


Figure 5. September 1984: (a) areas where herring larvae abundance exceeded $100 \cdot m^{-2}$; (b) areas where abundance of big herring larvae (>10 mm total length) exceeded $100 \cdot m^{-2}$.

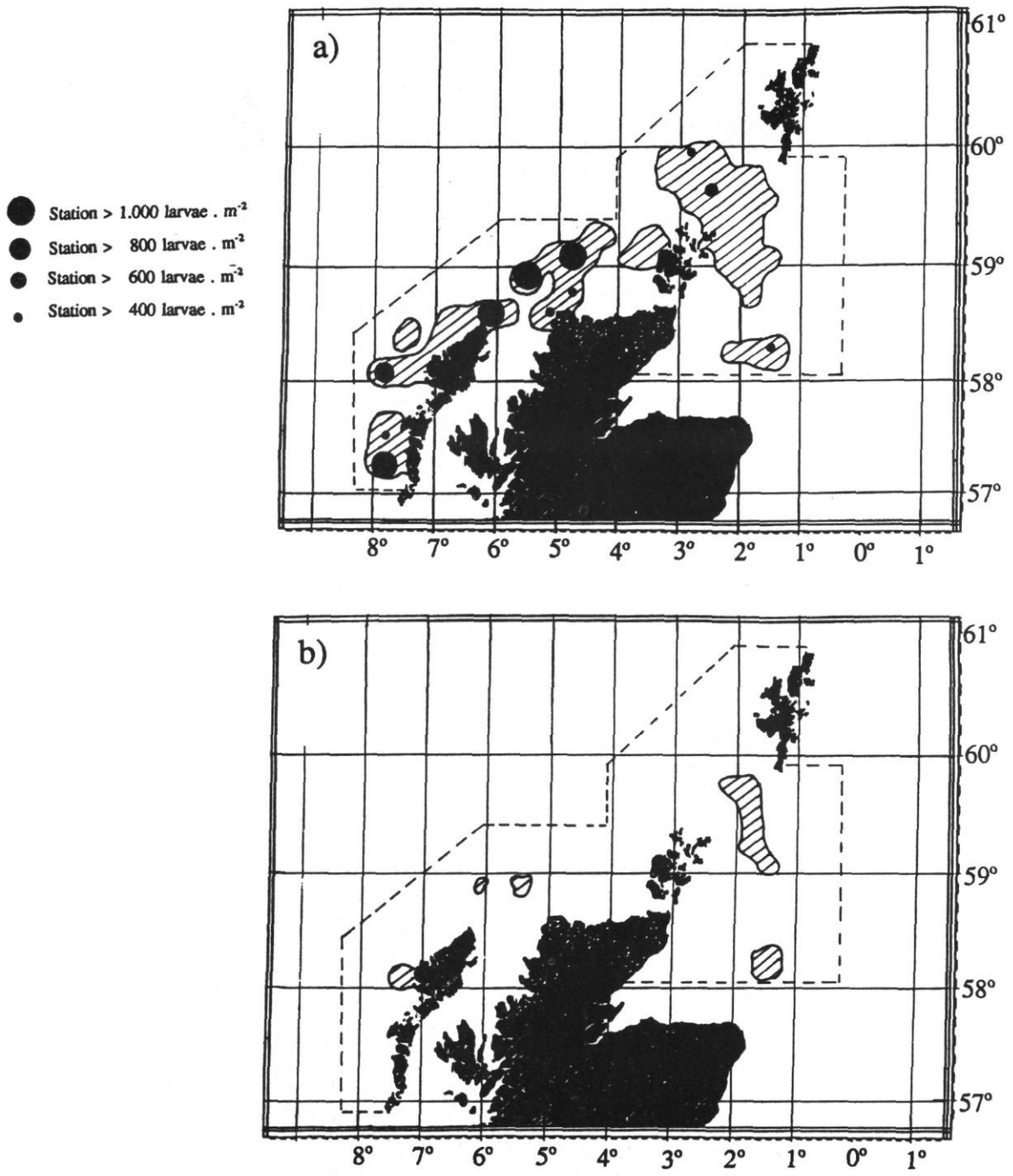
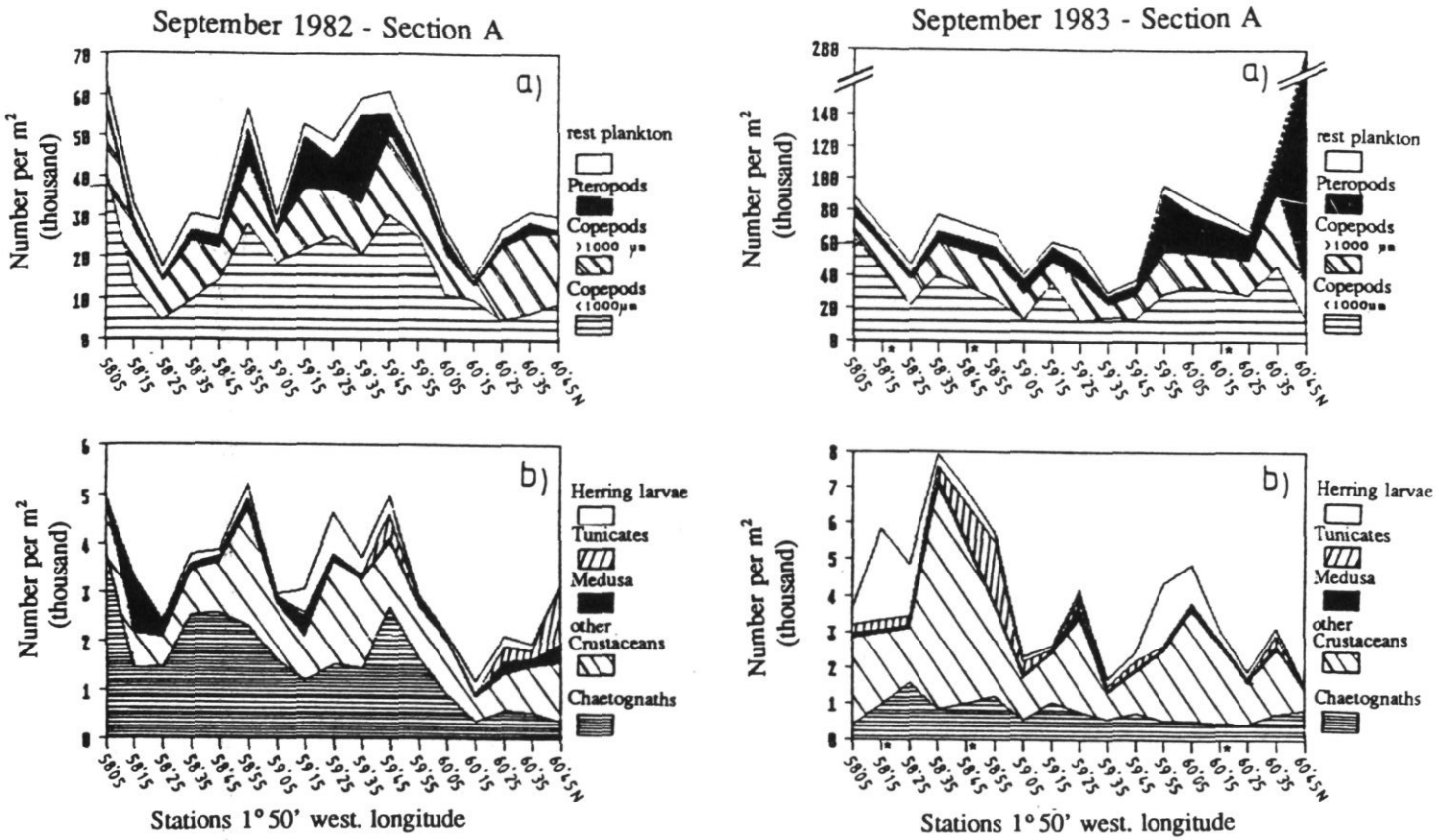


Figure 6. September 1985: (a) areas where herring larvae abundance exceeded $100 \cdot m^{-2}$; (b) areas where abundance of big herring larvae (>10 mm total length) exceeded $100 \cdot m^{-2}$.


September 1982 - Section A
September 1983 - Section A
a)
b)
Number per m² (thousand)
rest plankton
Pteropods
Copepods >1000 µm
Copepods <1000µm
Herring larvae
Tunicates
Medusa
other Crustaceans
Chaetognaths
Stations 1° 50' west. longitude

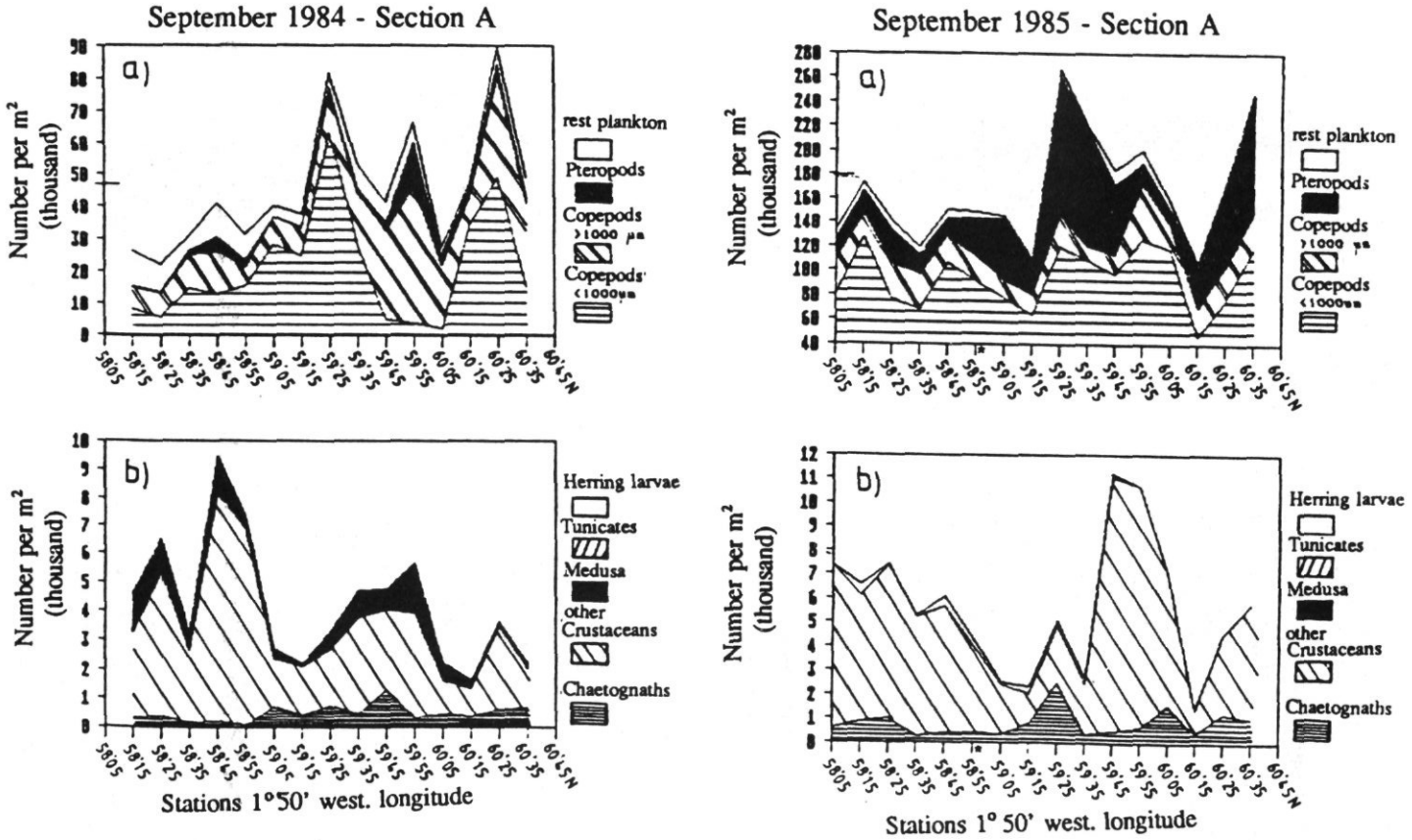
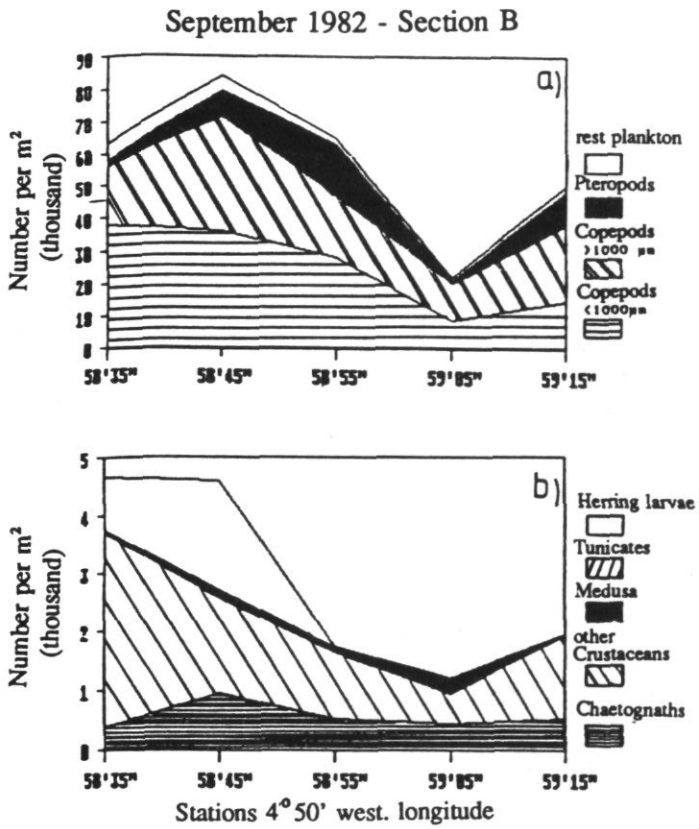
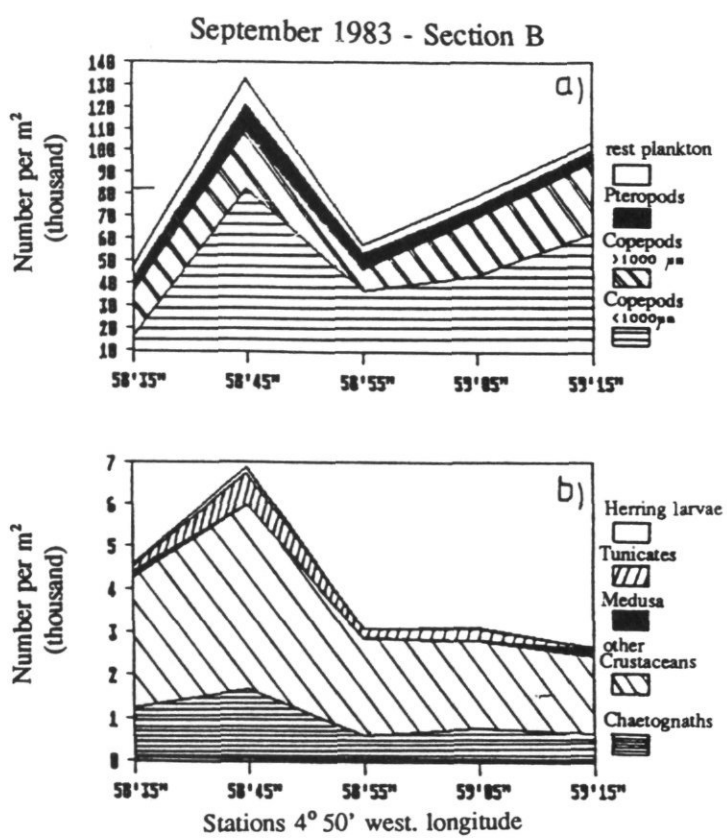


Figure 7. Abundance of zooplankton organisms of the various taxonomical groups as classified in Table 2 in section A in the years 1982 to 1985.


September 1982 - Section B
a)
Number per m² (thousand)
rest plankton
Pteropods
Copepods >1000 µm
Copepods <1000µm
58'35" 58'45" 58'55" 59'05" 59'15"
b)
Herring larvae
Tunicates
Medusa
other Crustaceans
Chaetognaths
Stations 4° 50' west. longitude


September 1983 - Section B
a)
Number per m² (thousand)
rest plankton
Pteropods
Copepods >1000 µm
Copepods <1000µm
58'35" 58'45" 58'55" 59'05" 59'15"
b)
Herring larvae
Tunicates
Medusa
other Crustaceans
Chaetognaths
Stations 4° 50' west. longitude

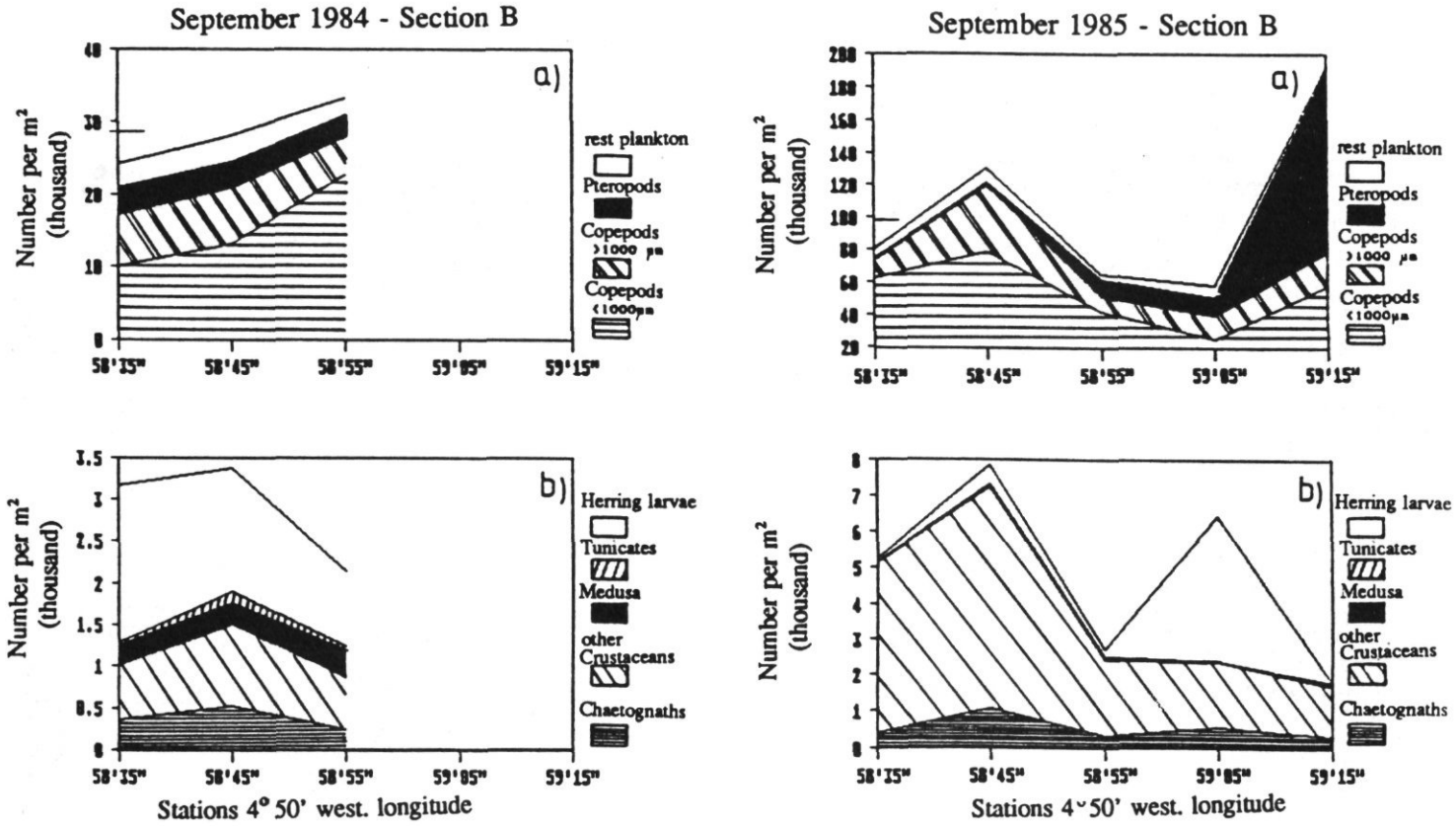
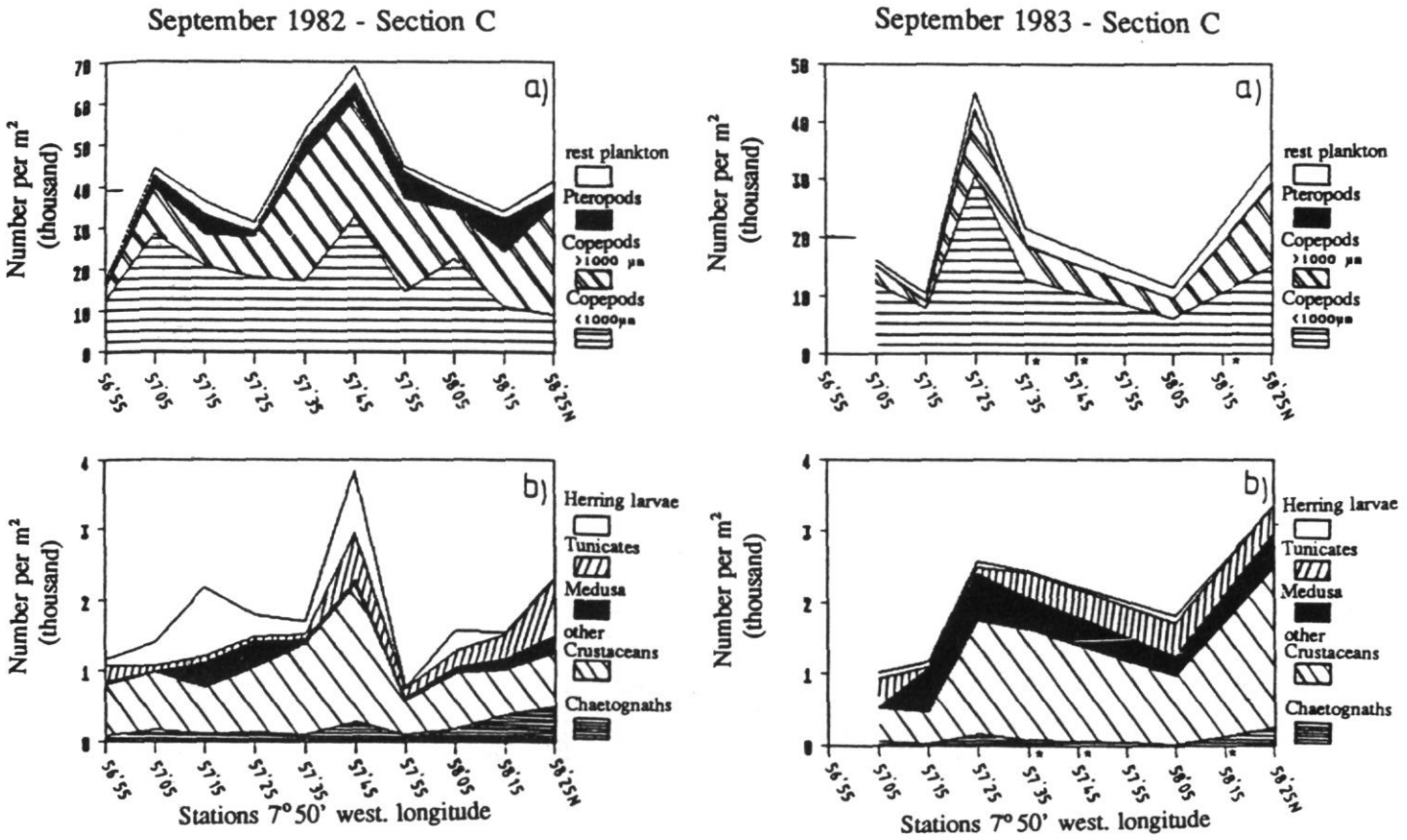


Figure 8. Abundance of zooplankton organisms of the various taxonomical groups as classified in Table 2 in section B in the years 1982 to 1985.


September 1982 - Section C
September 1983 - Section C
Number per m² (thousand)
a)
b)
rest plankton
Pteropods
Copepods >1000 µm
Copepods <1000µm
Herring larvae
Tunicates
Medusa
other Crustaceans
Chaetognaths
56°55
57°05
57°15
57°25
57°35
57°45
57°55
58°05
58°15
58°25N
Stations 7° 50' west. longitude

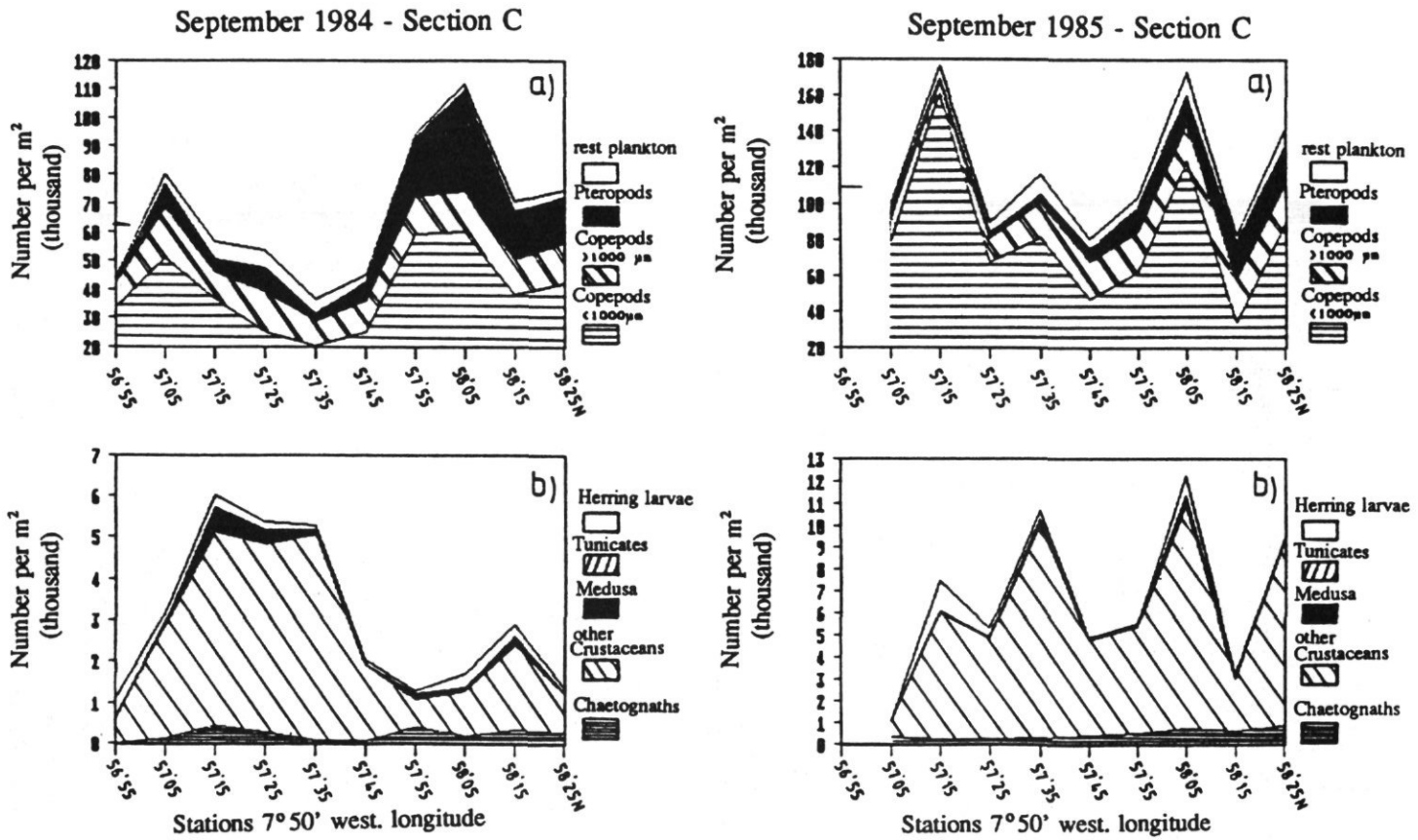


Figure 9. Abundance of zooplankton organisms of the various taxonomical groups as classified in Table 2 in section C in the years 1982 to 1985.

Table 1. Presence of taxa expressed as percentage of stations on which the specific taxon has been found; n = number of analysed stations.

			1982		1983		1984		1985	
		Taxon	n 37	% 100	n 30	% 100	n 35	% 100	n 33	% 100
Chaetognatha		Chaetognatha Rest	7	19	8	27	17	49	2	6
		Sagitta sp.	37	100	30	100	35	100	33	100
Hydrozoa	Hydroida	Medusa	32	86	24	80	27	77	17	52
		Siphonophora	26	70	10	33	33	94	16	48
Ctenophora		Ctenophora	0	0	7	23	8	23	6	18
Tunicata	Larvacea	*Oikopleura* sp.	4	11	4	13	8	23	2	6
	Thaliacea	*Salpa fusiformis*	0	0	28	93	20	57	1	3
	Thaliacea	*Doliolum* sp.	27	73	23	77	15	43	12	36
Phyllopoda	Cladocera	*Podon* sp.	23	62	28	93	32	91	32	97
	Cladocera	*Evadne spinifera*	26	70	29	97	31	89	30	91
	Cladocera	*Evadne nordmanni*	23	62	29	97	24	69	30	94
Eucarida		Euphausiacea	37	100	25	83	35	100	31	94
		Zoea/Decapod larvae	37	100	29	97	35	100	33	100
Peracarida		Mysidacea	6	16	15	50	20	57	9	27
	Amphipoda	Hyperiidae	37	100	30	100	34	97	32	97
	Amphipoda	Gammaridae	0	0	0	0	0	0	0	0
		Cumacea	3	8	0	0	5	14	2	6
		Isopoda	5	14	0	0	9	26	3	9
Copepoda	Calanoida	*Temora* sp.	37	100	30	100	35	100	33	100
	Calanoida	*Acartia* sp.	37	100	30	100	35	100	33	100
	Calanoida	*Para/Pseudocalanus*	37	100	30	100	35	100	33	100
	Cyclopoida	*Oithona* sp.	37	100	29	97	32	91	33	100
	Calanoida	*Calanus finmarchicus*	37	100	30	100	35	100	33	100
	Calanoida	*C. finmarchicus* cop.	36	97	25	83	35	100	33	100
	Calanoida	*Eucalanus* sp.	12	32	18	60	0	0	2	6
	Calanoida	*Centropages* sp.	37	100	30	100	35	100	33	100
	Calanoida	*Metridia* sp.	37	100	27	90	35	100	33	100
	Calanoida	*Candacia armata*	37	100	30	100	34	97	33	100
	Calanoida	*Anomalocera patersoni*	4	11	3	10	6	17	4	12
	Calanoida	*Rhincalanus nasatus*	29	78	24	80	11	31	16	48
		Copepod nauplii	22	59	27	90	10	29	16	48
Cirripedia		Balanid nauplii	11	30	9	30	5	14	9	27
Pteropoda	Thecosomata	*Limacina retroversa*	37	100	30	100	35	100	33	100
	Gymnosomata	*Clione limacina*	23	62	25	83	4	11	20	61
Polychaeta	Errantia	*Tomopteris* sp.	21	57	24	80	20	57	8	24
Tentaculata	Bryozoa	Cyphonautes larvae	15	41	23	77	24	69	28	85
Teleostei		Fish larvae	31	84	27	90	9	26	33	100
		Herring larvae	37	100	29	97	35	100	29	88

groups could not be separated at the 50% distance level at all, as the zooplankton composition was so homogeneous. The level of "significance" was therefore assumed at 45%, where two more important branches exist in the dendogram plus a third one, which, however, represents only one station. Usually the 30 stations are grouped in three main clusters, each group largely corresponding with the geographical positions of sections A, B and C (Fig. 14). The internal resemblance of the western and eastern clusters (sections A and C) was less than the resemblance between one of these clusters and the cluster in between, i.e. the Cape Wrath assemblage, where no one distinct group could be identified at all in 1985 (Fig. 14d).

The representative plankton communities which were determined for each cluster in each year (Table 3) always included more or less the same taxa. Perfect indicators, i.e. species that occur in all samples of one group and in no others (Field *et al.*, 1982), were lacking.

In 1985 the zooplankton was not only more homogeneous in the whole area, its abundance per m^2 was very high as well. Exceptions were "Atlantic species" such as salps, siphonophora, and appendicularia.

To get an idea of the variability of the zooplankton community in space and time, cluster analysis of the stations which had been sampled in the four years was carried out. When the cut-off level for defining clusters is set at 57%, three different groups result (Figs. 15a–c).

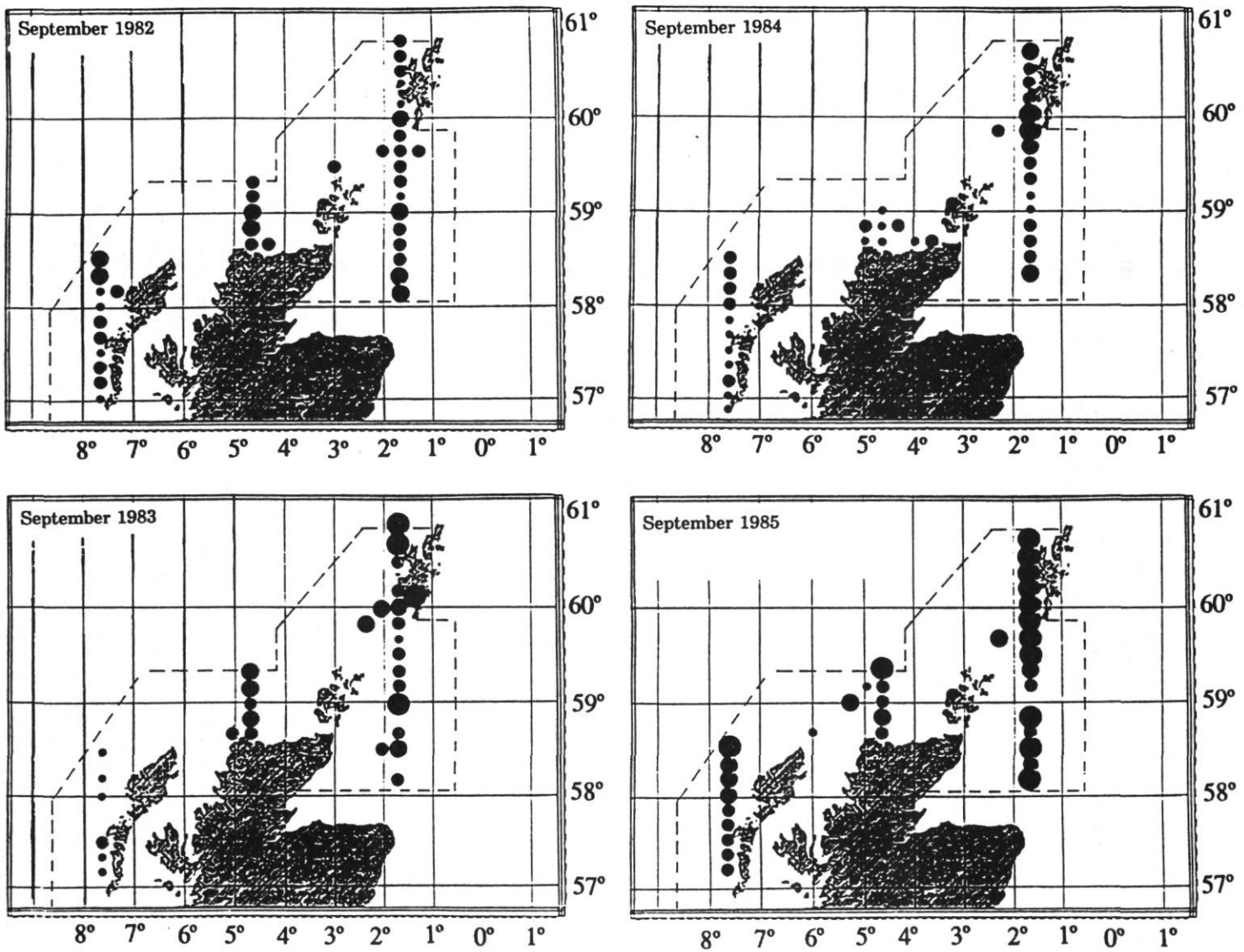
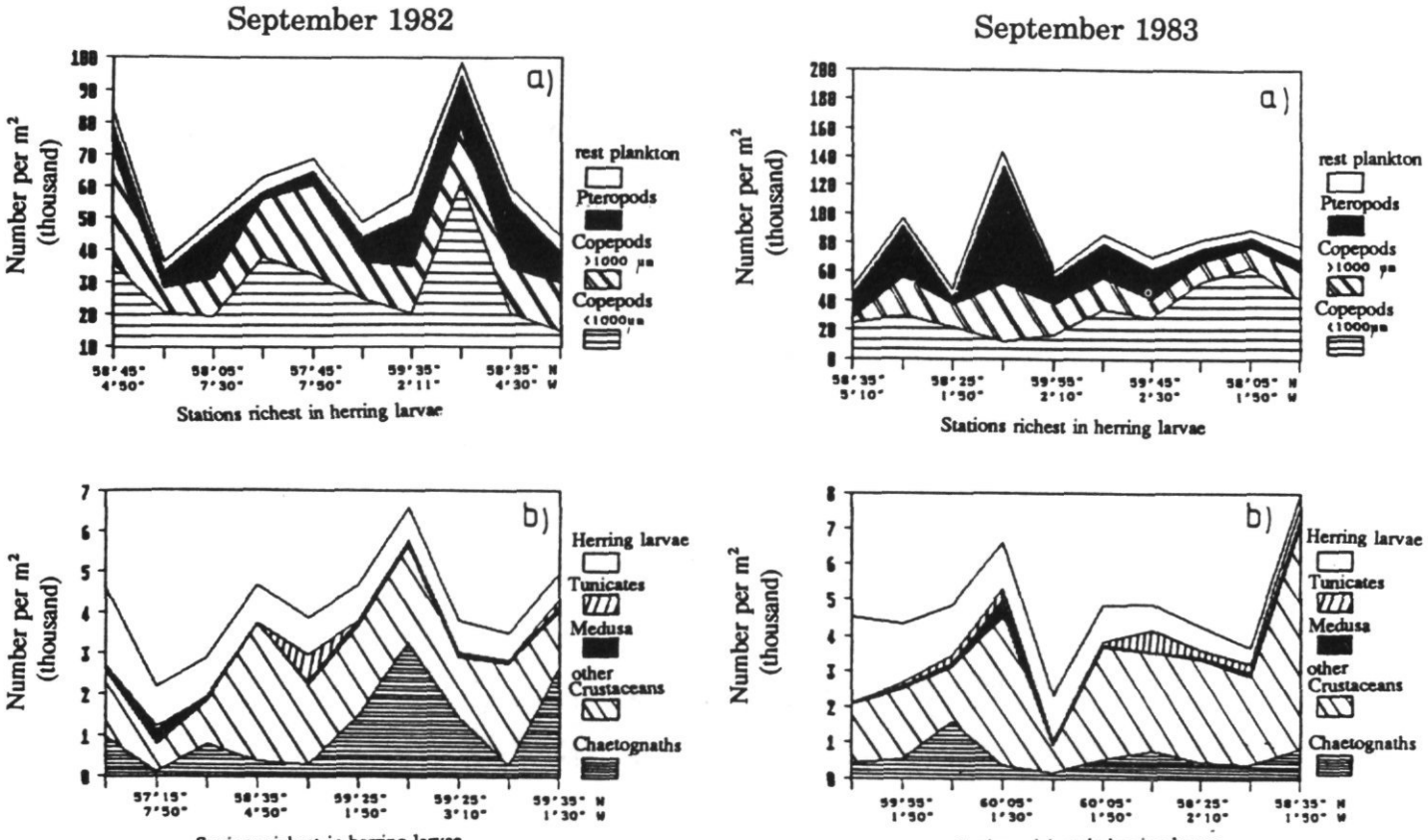


Figure 10. Distribution of zooplankton biomass, expressed as grams dry weight per m^2.


September 1982
September 1983
a)
b)
Number per m² (thousand)
rest plankton
Pteropods
Copepods >1000 µm
Copepods <1000µm
Herring larvae
Tunicates
Medusa
other Crustaceans
Chaetognaths
Stations richest in herring larvae

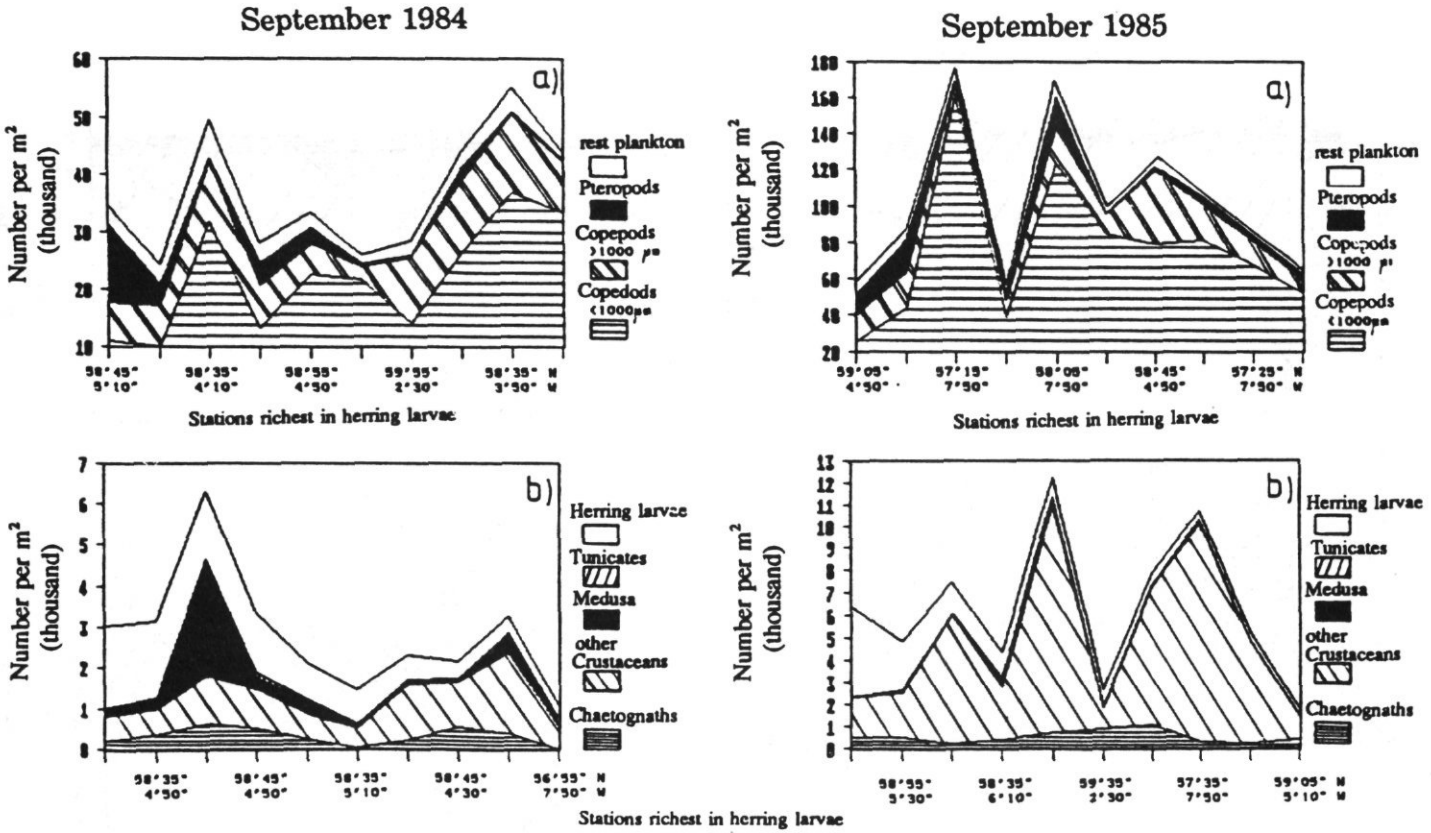


Figure 11. Abundance of zooplankton organisms of the various taxonomical groups as classified in Table 2 in the first ten stations richest in herring larvae.

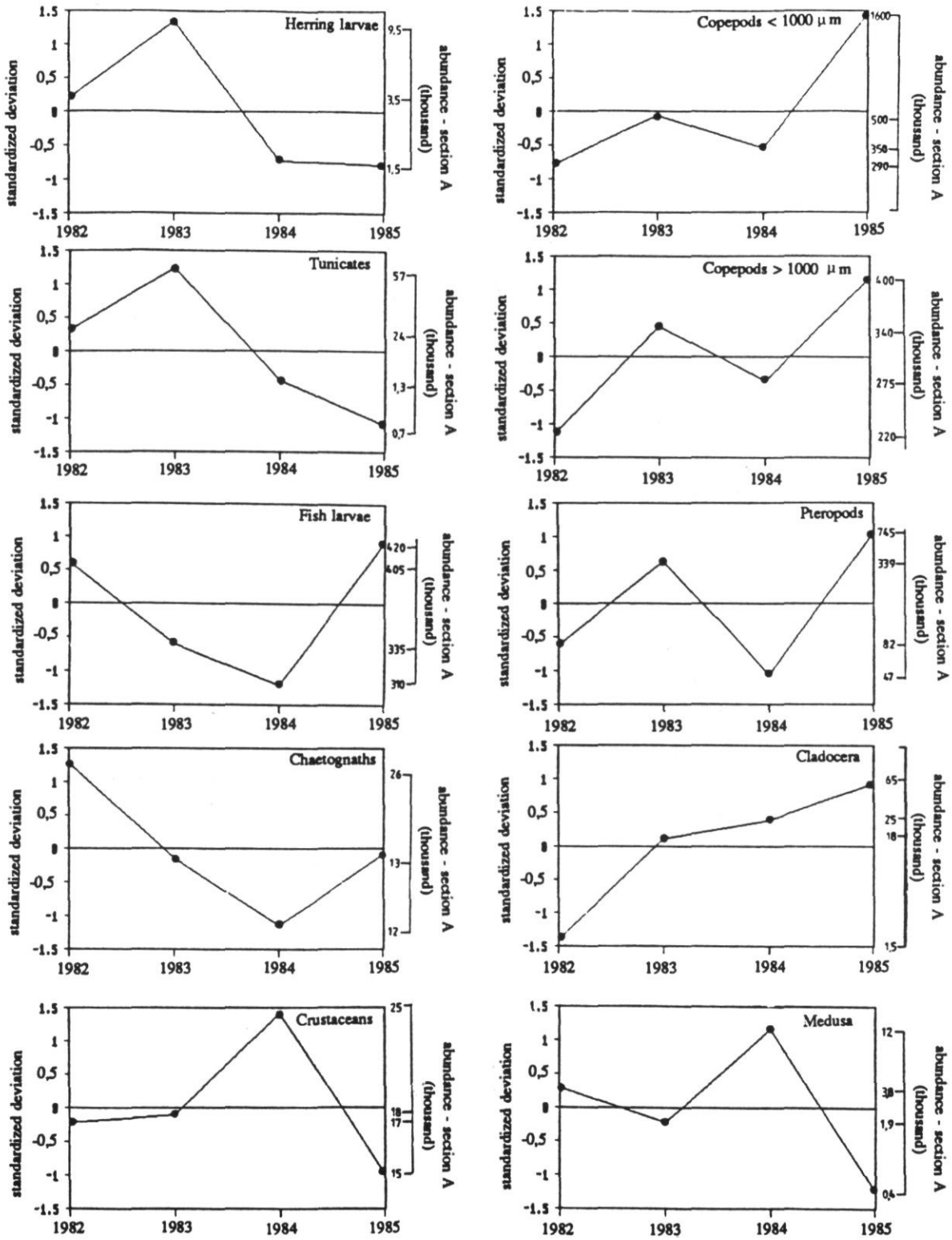


Figure 12. Year-to-year fluctuations in the abundance of some zooplankton taxa in section A (eastern area).

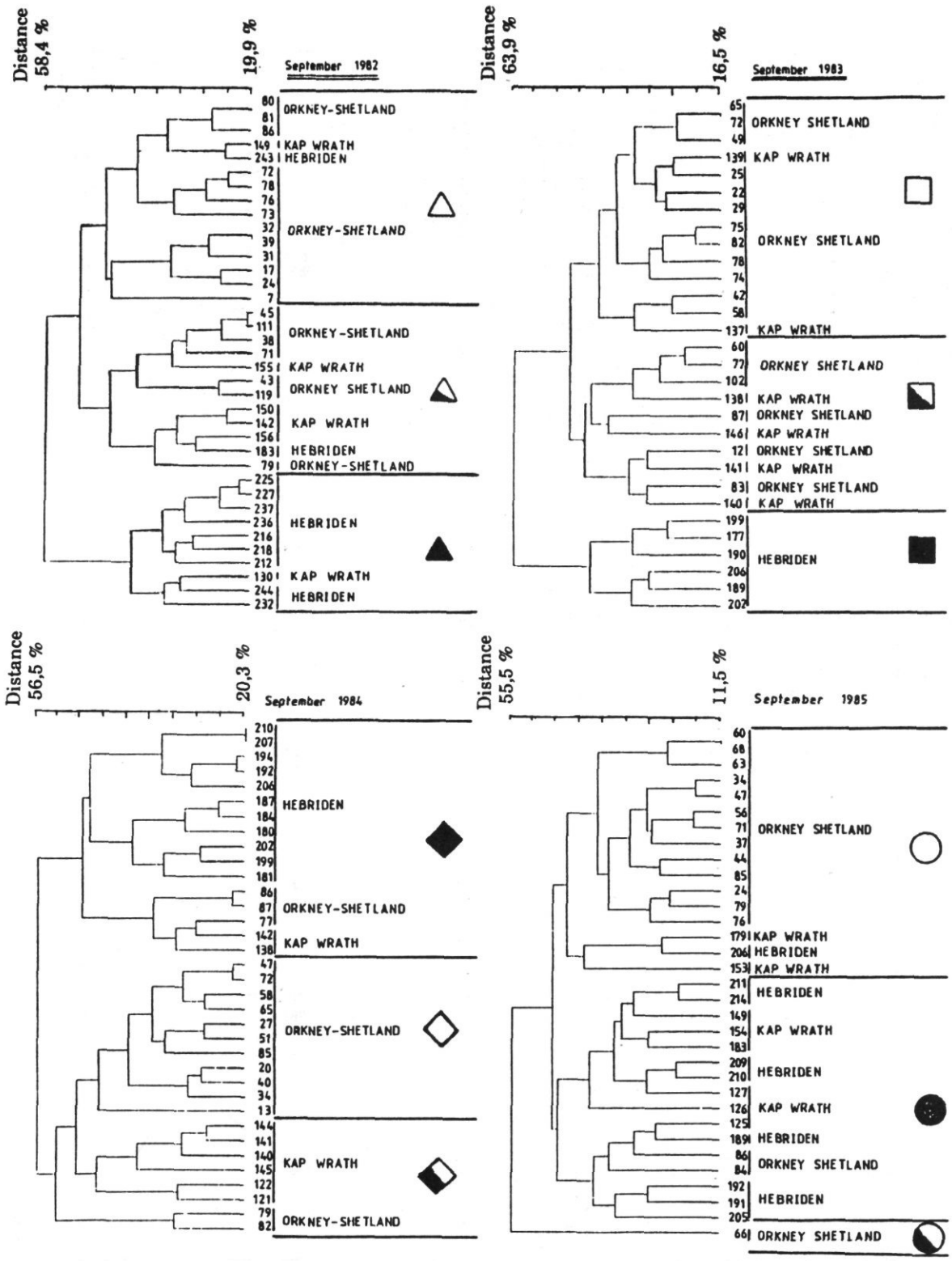


Figure 13. Dendograms showing classification of stations analysed for zooplankton composition in ICES areas IVa and VIa in 1982, 1983, 1984, and 1985. The symbols correspond to those shown in Figure 14.

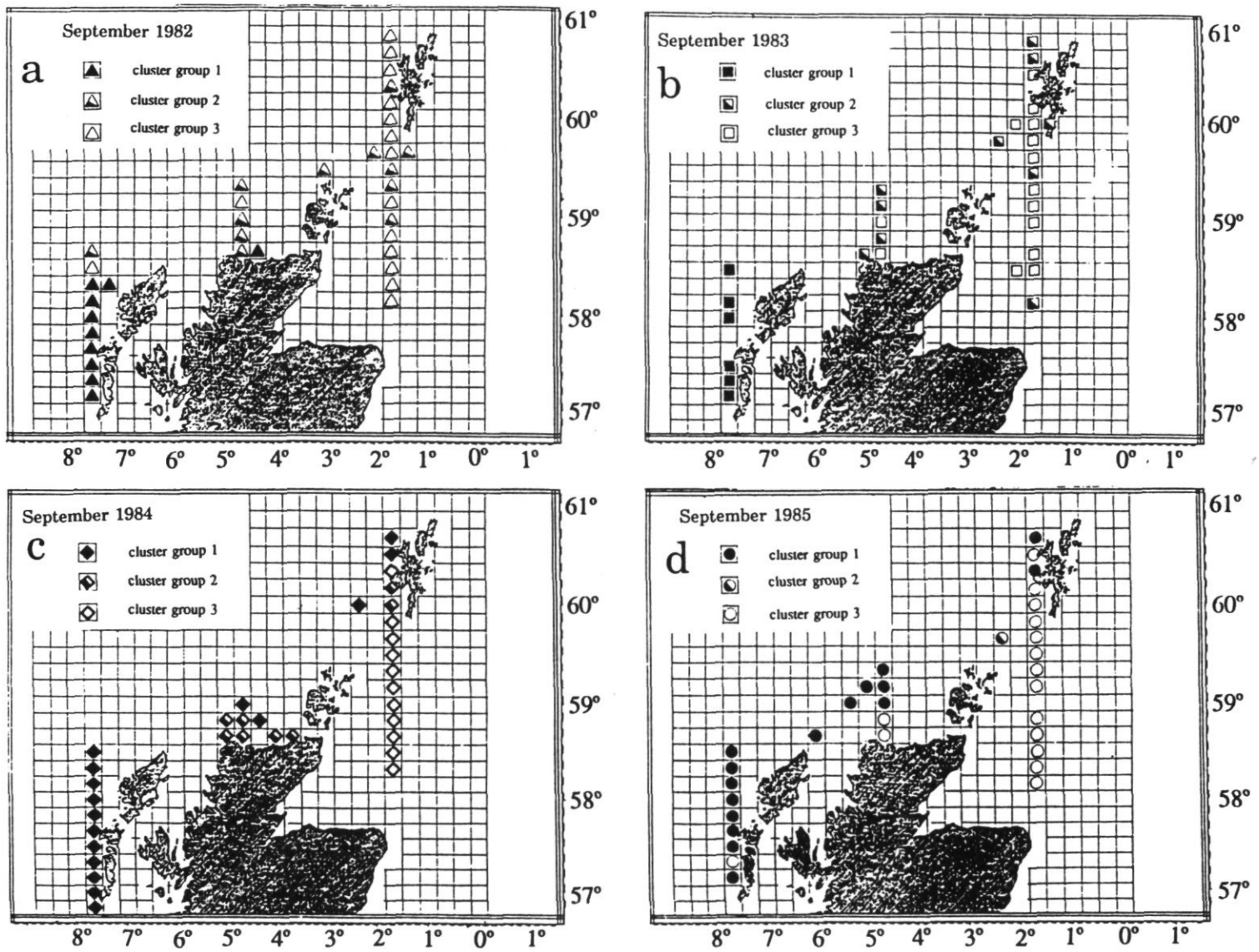


Figure 14. Geographical distribution of the zooplankton cluster groups shown in Figure 13.

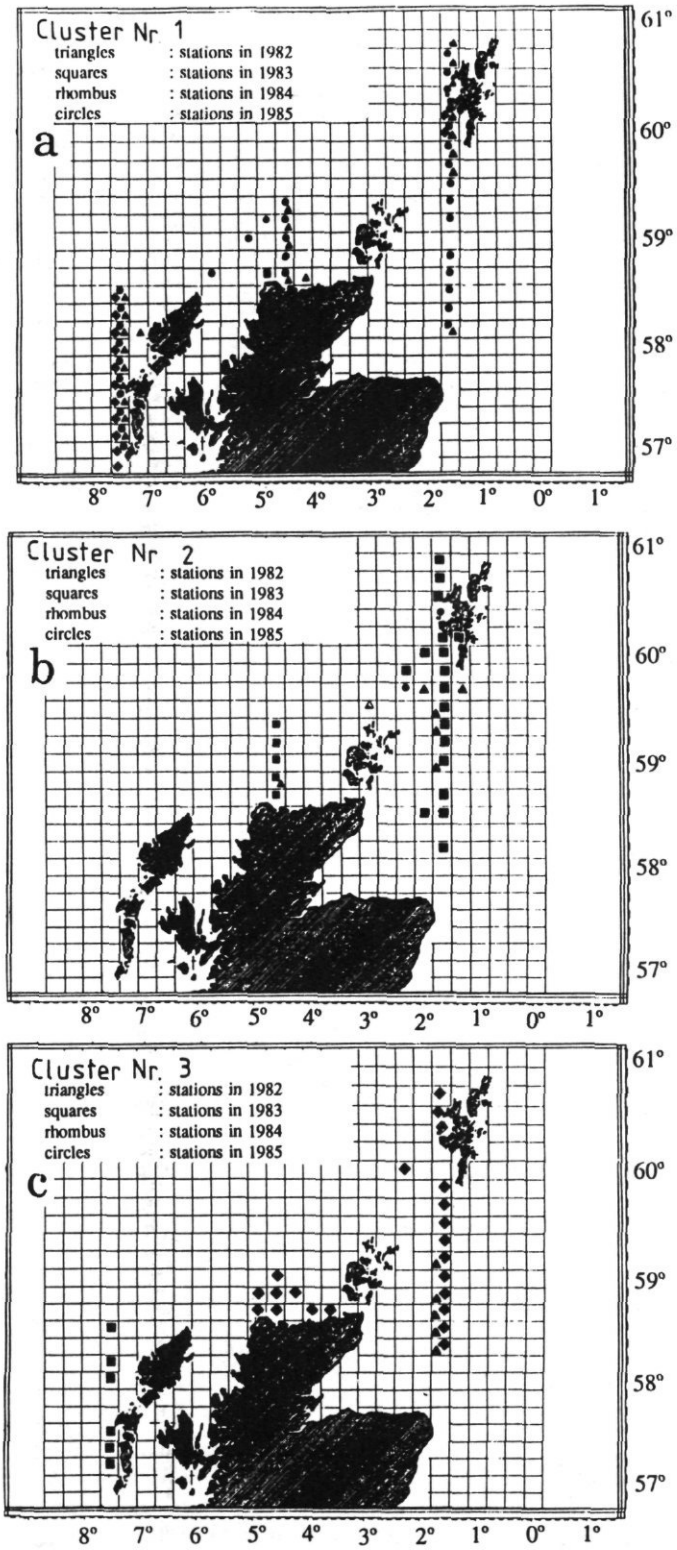


Figure 15. Geographical distribution of the three clusters identified when all stations of all years were clustered and the measure of similarity is set at 57%.

Table 2. Taxonomical composition of zooplankton groups as used to describe variation of taxa abundance in space and time.

Chaetognata	*Sagitta* spp.
	Other Chaetognaths
Medusa	Siphonophora
	Other Medusas
	Ctenophora
Tunicata	*Oikopleura* sp.
	Salpa fusiformis
	Doliolum gegenbauri
Copepoda <1000 m	*Temora* sp.
	Acartia sp.
	Para/Pseudocalanus
	Oithona sp.
Copepoda >1000 m	*Calanus finmarchicus* ad + v. cop
	Calanus finmarchicus cop.
	Eucalanus sp.
	Centropages sp.
	Candacia armata
	Anomalocera patersoni
	Rhincalanus nasatus
Other crustaceans	*Podon* sp.
	Evadne spinifera
	Evadne nordmanni
	Euphausiacea (*Thyssanoessa* sp.)
	Zoea/Decapod larvae
	Mysidacea (*Gastrosaccus* sp.)
	Hyperiidae
	Gammaridae
	Cumacea
	Isopoda
Pteropoda	*Limacina retroversa*
	Other Pteropods
	(mainly *Clione limacina*)
Herring larvae	

The Hebrides area is the most conservative one, as for three years the identical group, cluster no. 1, was found (Fig. 15a). Only in 1983 was the plankton community west of the Hebrides any different, described by cluster no. 3 (Fig. 15c). The other two areas were more variable between years. In 1985, when a homogeneous distribution of zooplankton was observed, one group, cluster no. 1, characterized all areas. In 1982 this plankton composition was also found at some stations in the Cape Wrath and Orkney–Shetland Islands area, characterizing mainly the western part of the research area only. The other two areas were characterized by cluster no. 2 mainly in 1983 and to a lesser extent in 1982. Cluster no. 3 was found there in 1984 and at a very few stations in 1982, too.

Discussion

Biological surveys may produce complex data, depending on the diversity of species collected, the accompanying abiotic environmental data, and the methods of analysis. In principle, patterns and relationships can be extracted from such data, but the question still largely unsolved is how and whether this will result in a substantial insight into ecological functions, alterations, or even structures. Ecological research often suffers from the heterogeneity of ecosystems, but in addition comparative studies of the results may suffer severely from differences in the precision of the material analysis as well as in the methods applied for processing the data. The interpretation of results depends to a large extent on experience, intuition, and ingenuity. Statistical treatment of ecological data as well as their modelling often becomes difficult and unreliable because of the lack of time series and routine measurements. Ecologists are therefore often confronted with the problem of coping with many confounding factors.

This investigation shows the variability of zooplankton abundance, distribution, and composition in a relatively small sea area influenced and altered by the North Atlantic Current and advection of different water masses.

One clear, but by no means new result, is that the areal extent of the spawning ground of the autumn spawning herring is relatively fixed, though patchy and changeable from one year to the next. Bigger herring larvae are far less abundant in the Hebrides area and a larvae drift to the Fair Isle area can be expected.

The zooplankton in the whole area is permanently dominated by three species of small herbivorous copepods and by the large *Calanus finmarchicus* and *Metridia* sp. The abundance of both groups is correlated. The herbivorous pteropod *Limacina retroversa* is another dominating form, but its distribution and abundance in space and time were found to be more variable and patchy. When biomass in dry weight is taken into account, big crustaceans, such as euphausids, mysids, Hyperids and decapod larvae, become important too. But these groups never exceeded 9% of the total biomass expressed in dry weight, sometimes even less than 5%, whereas the two large copepods (*Calanus* and *Metridia*) represented 60% to almost 75% of zooplankton dry weight in all four years.

With regard to the average abundance of copepods and pteropods, the extremes were found at about $40 \cdot 10^3$ specimens per m^{-2} at minimum and $180 \cdot 10^3$ specimens per m^{-2} at maximum in the two eastern areas (Cape Wrath and the Orkneys–Shetland Islands). Both areas were richer in these zooplankton animals in 1983 and 1985 than in 1982 and 1984. The extreme values in the Hebrides area were lower, amounting to $20 \cdot 10^3$ specimens per m^{-2} at minimum and to $110 \cdot 10^3$ specimens per m^{-2} at maximum. The highest zooplankton concentrations west of the Hebrides were measured in 1984 and 1985. These findings correspond widely with the biomass measurements (Table 4). The average amount of total zooplankton measured as ash-free dry weight was higher in the three areas in 1985 and distinctly lower in the other years. Usually, average zooplankton biomass was lower in the west.

Table 3. Representative plankton communities, i.e. arithmetic means of abundance ($n \cdot m^{-2}$), of the clusters shown in Figure 13. The symbols plus and minus indicate whether the difference in abundance of the specific taxon was large or small compared to the neighbouring community.

	▲	◭	△		■	◪	□
	cluster group 1	cluster group 2	cluster group 3		cluster group 1	cluster group 2	cluster group 3
1982	RPG	RPG	RPG	1983	RPG	RPG	RPG
Chaetognatha	19−−	70	78+	Chaetognatha	14−−	101++	44
Sagitta sp.	202−−	1219+	1409++	*Sagitta* sp.	97−−	635+	708++
Medusa	96−	127	183++	Medusa	82	142+	65−
Siphonophora	20	11−	41+	Siphonophora	331++	15	5−−
Ctenophora	0	0	0	Ctenophora	7	0	10
Oikopleura sp.	38++	0	0	*Oikopleura* sp.	3−	9+	5
Salpa fusiformis	0	0	0	*Salpa fusiformis*	247++	174	167+
Doliolum geg.	123	30−	185+	*Doliolum geg.*	29−	91	123++
Podon sp.	115++	16−	40	*Podon* sp.	434++	182	104−
Evadne spinifera	84+	39−	49	*Evadne spinifera*	207	202−	312+
Evadne nordmanni	120++	20−	28	*Evadne nordmanni*	224−	505	559+
Euphausiacea	358	571++	179−	Euphausiacea	76−	196+	130
Zoea/Decapoda	360+	333	210−	Zoea/Decapoda	184−	257+	283
Mysidacea	3	6	3	Mysidacea	7−	197+	22
Hyperiidae	70−−	596++	480+	Hyperiidae	67−−	809++	711
Gammaridae	0	0	0	Gammaridae	0	0	0
Cumacea	5	1	1	Cumacea	0	0	0
Isopoda	2	8	0	Isopoda	0	0	0
Temora sp.	1285−−	8462++	7947+	*Temora* sp.	1633−−	11331++	10915+
Acartia sp.	5280+	4295	2896−	*Acartia* sp.	2814−−	10295++	4792+
Para/Pseudocalanus	13963+	12713+	4238−	*Para/Pseudocalanus*	924−−	17736++	11620+
Oithona sp.	183+	163	100−−	*Oithona* sp.	262	722+	200+
Calanus finmarchicus	4853−−	6253+	7922+	*Calanus finmarchicus*	2226−−	11974++	9380+
C. finmarchicus cop.	4949++	2841	1050−−	*C. finmarchicus* cop.	773−−	4943+	3594+
Eucalanus sp.	2−	17+	7	*Eucalanus* sp.	34−	74+	41
Centropages sp.	2909+	2846+	1371−	*Centropages* sp.	280−−	3444++	2478+
Metridia sp.	1890−	2818+	2257	*Metridia* sp.	2861	3030+	1465−
Candacia armata	344−	1425+	450	*Candacia armata*	72−−	396++	241
Anomalocera pat.	1	1	3	*Anomalocera pat.*	0	5	0
Rhincalanus nas.	7−−	226	389++	*Rhincalanus nas.*	44−−	294+	113
Copepod nauplii	92+	51	45−	Copepod nauplii	35−	228+	66
Balanid nauplii	87++	0−	28	Balanid nauplii	0−	13	88++
Limacina retroversa	6436+	9031+	4033−	*Limacina retroversa*	253−−	40800++	11536+
Clione limacina	2−	54+	37	*Clione limacina*	16−	306++	135
Tomopteris sp.	17	15−	27+	*Tomopteris* sp.	22−	28	50+
Cyphonautes	15	8−	35+	Cyphonautes	18−	97+	71
Fish larvae	14−	19	19	Fish larvae	11−	14	20+
Herring larvae	465+	580+	120−	Herring larvae	64−−	496+	536++

The overall variability in zooplankton biomass in the area is regarded as being fairly low. In three years the extremes did not differ by more than a factor of 3, and when the plankton-rich year of 1985 is taken into consideration the factor is 5.6 for four years. The number of copepod and pteropod specimens per m^2, as shown in Figures 7–9, differed on average by a factor of 4 in the first three years and by a factor of 9 when 1985 is included. These findings may imply a relatively stable ecological situation in the whole herring spawning area.

The fluctuation of certain zooplankton taxa indicate, however, that the environmental conditions differ from year to year. The co-occurrence of high and low numbers of tunicates and herring larvae respectively on the eastern stations indicates that the Fair Isle region was influenced to a different degree by water masses of southern and oceanic origin. Salps, the copepod *Rhincalanus nasatus*, as well as some medusa are regarded as such indicator organisms (Fraser, 1965), and all were much more abundant in 1982 and 1983 (Table 5) than in the other two years.

If these taxa had drifted from west to east the number

Table 3 *cont.*

1984	◆ cluster group 1 RPG	◈ cluster group 2 RPG	◇ cluster group 3 RPG
Chaetognatha	55	92++	12−
Sagitta sp.	246−	285	497+
Medusa	39−	40	41+
Siphonophora	113−−	772++	700+
Ctenophora	7	14+	4−
Oikopleura sp.	33++	7	0−
Salpa fusiformis	34	22−	42+
Doliolum geg.	11	8−	11
Podon sp.	413+	273−−	683++
Evadne spinifera	446+	131−−	654++
Evadne nordmanni	280+	74−−	629++
Euphausiacea	167	136−−	371++
Zoea/Decapoda larvae	367	439+	253−
Mysidacea	7−	61+	21
Hyperiidae	350	198−−	795++
Gammaridae	0	0	0
Cumacea	0	5	3
Isopoda	0	7	1
Temora sp.	3890−	7626+	5701
Acartia sp.	8813++	3736	1100−−
Para/Pseudocalanus	21788++	4822−	15024
Oithona sp.	688+	125−	162
Calanus finmarchicus	6674−	8772+	8181
C. finmarchicus cop.	1605−	2135+	1841
Eucalanus sp.	0	0	0
Centropages sp.	1530+	1083	233−
Metridia sp.	3438++	1257	1014−−
Candacia armata	396	479+	121−−
Anomalocera pat.	2	4	1
Rhincalanus nas.	29+	10	1−−
Copepod nauplii	61	25−	277+
Balanid nauplii	5	5	21
Limacina retroversa	8237+	5025+	1622−
Clione limacina	3	3	0
Tomopteris sp.	6−	16+	8
Cyphonautes	41+	28	21−
Fish larvae	5−	22+	11
Herring larvae	215+	1031++	127−

1985	● cluster group 1 RPG	◔ cluster group 2 RPG	○ cluster group 3 RPG
Chaetognatha	3	0	0
Sagitta sp.	509−−	919+	797+
Medusa	29	53+	22−
Siphonophora	81++	0−−	8
Ctenophora	8+	0−	4
Oikopleura sp.	0	0	29+
Salpa fusiformis	0	11	0
Doliolum geg.	21+	0−	5
Podon sp.	2450+	363−	3509++
Evadne spinifera	130	0−	234+
Evadne nordmanni	168	0−	385+
Euphausiacea	406+	107−	162
Zoea/Decapoda larvae	538	36−	612+
Mysidacea	8	0−	32+
Hyperiidae	145−	438+	229
Gammaridae	0	0	0
Cumacea	1	0	4
Isopoda	0	11	5
Temora sp.	6594−−	12179+	29872++
Acartia sp.	3694−−	9295+	7683
Para/Pseudocalanus	52358−	62286++	57729+
Oithona sp.	591−	962	1230++
Calanus finmarchicus	9791	5021−	11784++
C. finmarchicus cop.	1521	962−−	2672+
Eucalanus sp.	1	0	0
Centropages sp.	2720	1282−	4327+
Metridia sp.	3417−	3606	3829+
Candacia armata	180	134−	305+
Anomalocera pat.	64+	0−	3
Rhincalanus nas.	31+	0−	25
Copepod nauplii	61+	53−	54
Balanid nauplii	0−	630++	91
Limacina retroversa	23905+	120−−	33806++
Clione limacina	35	21−	240+
Tomopteris sp.	5−	406++	8
Cyphonautes	51	0−	222+
Fish larvae	30+	11−	23
Herring larvae	608+	747+	215−

of herring larvae in the east would have been high as well because of a transport from the spawning grounds off the Hebrides and the north coast of Scotland. This was in fact the case (Figs. 7 and 12). The cluster analyses gave corresponding results: stations assembled in cluster no. 2, primarily typical for the Cape Wrath area, were exceptionally high in number also in the most eastern area in both 1982 and 1983.

Herring larvae drift through the Orkneys–Shetland Islands passage into the North Sea is assumed to be one of the important factors for high recruitment. As Atlantic influence favors this transport, the conditions in 1982 and 1983 should have been favorable. In contrast, the Atlantic influence seemed to be low in 1985, with very few of the indicator species found in the plankton samples.

As pteropods also originate in oceanic regions (Wiborg, 1955) their exceptionally high number, especially in the eastern area in 1985, is surprising. In this year *Limacina* was probably advected by water masses from the Norwegian Sea, i.e. from an area where the other, more southern indicator species, are not expected.

Table 4. Average amount of zooplankton (g m^{-2} ash-free dry weight) and standard deviation (%).

	1982	1983	1984	1985
Orkneys–Shetland Islands	1.46 ± 38%	1.97 ± 42%	1.69 ± 51%	3.63 ± 31%
Cape Wrath	1.70 ± 23%	1.73 ± 24%	0.65 ± 15%	2.00 ± 33%
Hebrides	1.13 ± 50%	1.01 ± 70%	0.83 ± 25%	2.00 ± 23%

Unfortunately, almost no hydrographical data were taken with the biological samples, as census of the herring larvae was the exclusive aim of the surveys. The only exceptions were surface temperature measurements taken in all four years, and the bottom temperature measurements in 1985 (Figs. 16 and 17). Surface temperatures between the western and eastern parts of the investigation area were very similar in September. The extremes did not deviate by much more than 1°C, and the temperatures at the surface were almost stable within the range 11.5°C to 12.5°C. In correspondence with the relatively homogeneous plankton distribution in 1985, the surface isotherms were less scattered in this year compared to 1982, 1983, and 1984. The bottom temperatures registered in 1985 increased from east to west, and, from the grouping of the near-bottom isotherms, it could be suggested that water was advected into the survey area from the east. But bottom temperature may simply have followed the depth contours (Fig. 18).

The sea area where herring larvae were found is, on the one hand, characterized by weak to absent stratification of the water column, as has already been described by Dietrich *et al.* (1975) for the waters framing the northern parts of the British Isles. On the other hand, extensions of gravel beds forming the bottom substratum in the Hebrides–Orkneys area are typical features of this spawning ground.

A fairly homogeneous zooplankton community indicated by the cluster analysis in 1985 and high zooplankton concentrations should favor recruitment. The advection of water masses from the east are more likely to be an unfavorable factor. As the situations were in contrast in 1985, it is questionable whether one has higher ecological importance.

It is very likely that neither situation explains the cause of recruitment success. The respective ICES working group reports say that VPA resulted in the following year-class strengths of North Sea herring: 1982: $13.5 \cdot 10^9$ recruits, 1983: $13.6 \cdot 10^9$ recruits, 1984: $22.7 \cdot 10^9$ recruits, 1985: $26.4 \cdot 10^9$ recruits. But recruits cannot be related to distinct stocks, and specific ecological situations such as are found in ICES areas VIa and VIa must not necessarily be correlated with recruit numbers of North Sea herring as a whole.

Table 5. Plankton indicators and their total abundance in the Orkneys–Shetland Islands area (Section A).

	1982	1983	1984	1985
Doliolum gegenbauri	2205	1722	135	75
Salpa fusiformis	–	2548	690	–
Rhincalanus nasatus	6001	3640	60	210
Medusa	2820	1344	630	315

Marine ecologists are asked for the function of biological systems. This implies an understanding of inorganic and organic matter and energy flux and its variability. The paths which matter and energy take in an ecosystem depends very much on species composition, relative abundance and size composition of species and on species biomass. All these factors are variable, not only resulting for instance in changing recruitment rates of fish stocks (Nellen, 1986) but in changing rates of sedimentation, consumption, and remineralization of organic matter, too. The processes of fish-stock recruitment and of long-term alterations in the function of the marine epipelagic ecosystem in relation to changing climate and anthropogenic loads are the most important aspects of marine biological science today.

Since 1967, much research vessel time has been invested in sampling plankton on important spawning grounds of North Sea herring just to count the herring larvae of successive seasons. Because of the high accuracy with which the hauls are made, a lot more information of major ecological importance could have been gained had countries agreed early on some additional analysis of the invertebrate plankton. This is especially true as ecological events in the specific sea area where the plankton is collected likely influence plankton and fish production of the North Sea and larger areas of the North Atlantic as well. The results would also have been valuable for comparing with the data from the plankton recorder program which is up to now the only effort being made to monitor plankton on a routine basis. Such programs have to be more intensive for the future understanding of biological–oceanographic processes.

Acknowledgment

We thank Dr Michael P. Sissenwine for critically commenting on the manuscript.

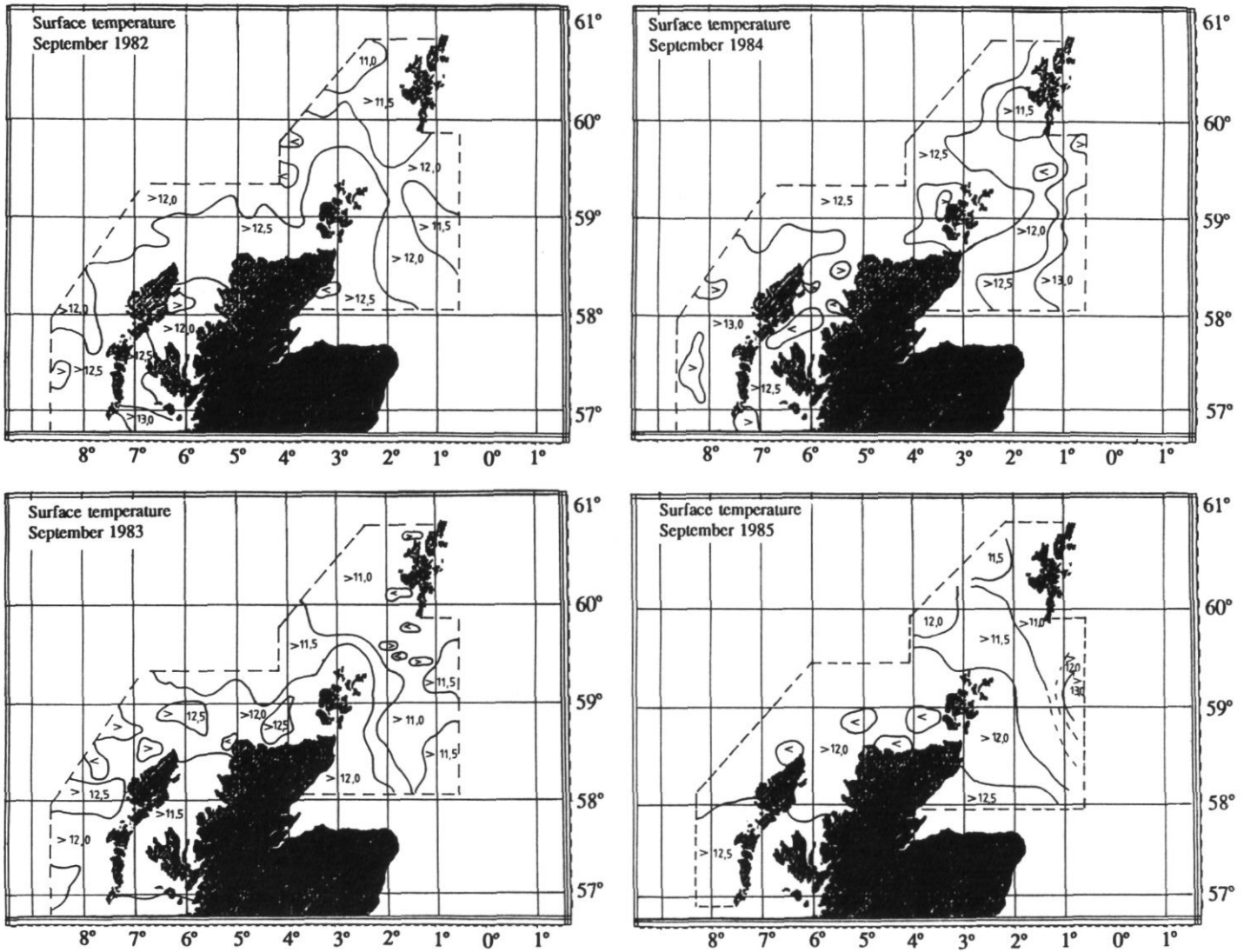


Figure 16. Distribution of surface temperatures in the area of investigation as measured during the survey in September 1982, 1983, 1984, and 1985.

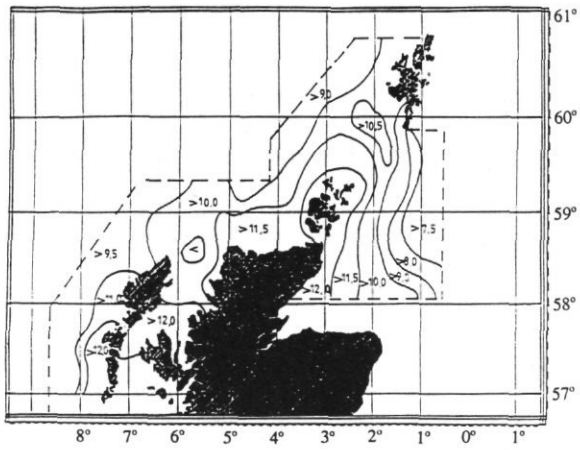


Figure 17. Bottom water temperature in September 1985.

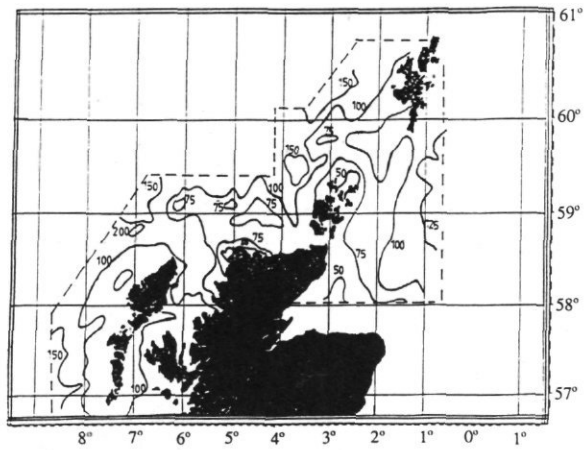


Figure 18. Water depth in meters.

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Climatic influences on the spawning success, growth, and recruitment of bass (*Dicentrarchus labrax* L.) in British waters

M. G. Pawson

Pawson, M. G. 1992. Climatic influences on the spawning success, growth, and recruitment of bass (*Dicentrarchus labrax* L.) in British waters. – ICES mar. Sci. Symp., 195: 388–392.

Sea bass (*Dicentrarchus labrax* L.) around Britain are near the northern limit of the species' geographical range, and climate changes have been implicated in historic fluctuations in year-class abundance. There is a major spawning and nursery area along the English coast of the central English Channel, where seasonal inshore water temperatures have undergone a cycle of warming, cooling, and warming again during the 1980s. This pattern is reflected in the abundance indices of juvenile bass, and their growth rate has also been positively related to temperature. It appears that the climate has had a strong influence on the recruitment of juvenile bass to the fishery in the English Channel, and that survival and growth are more dependent on temperature than on population density.

M. G. Pawson: Ministry of Agriculture, Fisheries and Food, Fisheries Laboratory, Lowestoft, Suffolk NR33 0HT, UK.

Introduction

The population of sea bass (*Dicentrarchus labrax* L.) in the southern North Sea and English Channel spawns during February–May in ICES Divisions VIIe and VIId (Thompson and Harrop, 1987). Its main nursery area in Division VIId is in the sheltered waters of the Solent and associated estuaries and harbours (Fig. 1). 0-group bass recruit there in July–August and remain inshore along the central southern English coast for 4–6 years, before recruiting to the adult population (Pawson *et al.*, 1987). Britain is near the northern limit of the geographical range of bass, and many authors have suggested important effects of sea temperature on distribution and growth (e.g. Holden and Williams, 1974; Kelley, 1988; Pawson *et al.*, 1987). Gonad maturation and spawning in bass appear to be regulated by temperature (Devauchelle, 1984) and it is thought that unusually low temperatures cause increased mortality of bass during their first winter (Kelley, 1988). It is therefore likely that sea temperatures before and during the spawning season might influence the distribution of spawning bass and the survival of larvae, and hence the level of recruitment to particular nursery areas, and that water temperatures there will influence the subsequent survival of juvenile bass. This paper examines the relationship between seasonal water temperatures and the abundance and growth of juvenile bass in the central English Channel during the 1980s.

Materials and methods

Year-class strength indices

Trawl surveys of juvenile (1- to 5-group) bass have been carried out over a standard grid of 32 stations in the Solent nursery area once or twice each year since 1981. Survey age-length-keys were prepared using length-stratified scale samples and, with the length frequency distribution of the catch of bass at each station, numbers at age per unit towing time were calculated (Pickett and Pawson, in prep.). On the assumption that the survey catchability of each age group was consistent throughout the survey period, and that mortality of juvenile bass did not change significantly through the 1980s, the mean abundance ($\bar{x}_a$) of y bass year classes caught at age a in the Solent nursery area was calculated using:

$$\bar{x}_a = \frac{1}{n_y} \sum_{n_y} x_{a,y}$$

where $x_{a,y}$=survey mean c.p.u.e. of year class y at age a, and n_y is the number of surveys in which bass year classes 1980–1989 were sampled at age a. The relative abundance ($\bar{x}_y$) of each year class is given by

$$\bar{x}_y = \frac{1}{n_a} \sum_{n_a} x_{a,y}/\bar{x}_a$$

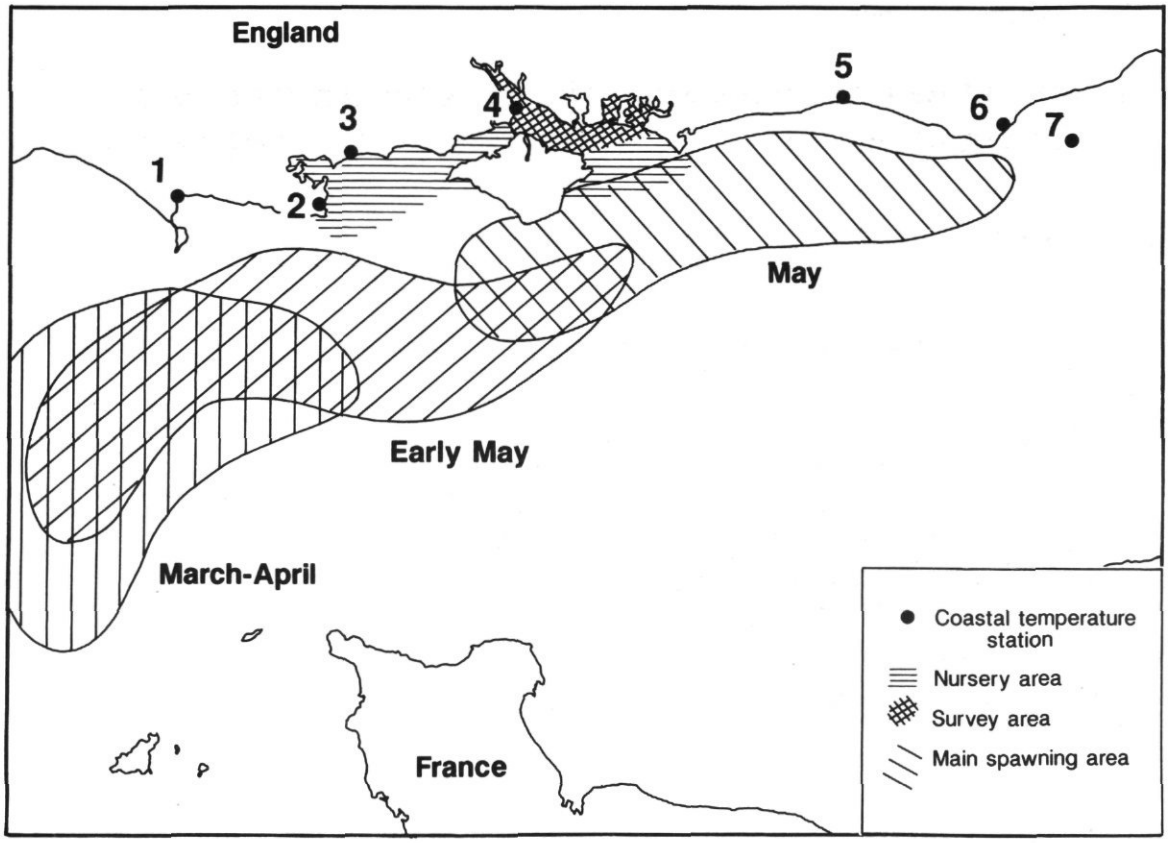

Figure 1. The central English Channel, showing bass spawning distributions, the main nursery area centred on the Solent, the trawl survey area, and coastal temperature recording stations (1 Weymouth; 2 Swanage; 3 Bournemouth; 4 Calshot/Fawley; 5 Shoreham; 6 Eastbourne; 7 Royal Sovereign light vessel).

where n_a is the number of surveys in which bass of year class y were sampled between 1980 and 1989.

Year-class growth

Survey age-length distributions were used to calculate mean lengths-at-age for each year class of pre-recruit bass. The length at the end of the third growing season (0- and 1-group bass were not well sampled in these surveys and the gear selectivity was maximal for 2- or 3-groups) for each of the year classes 1979–1988 was used as an index of growth.

Sea temperatures

Sea surface temperature data are available for several recording stations along the north coast of the central English Channel (Fig. 1) over most of the period November 1979 to November 1989. Using monthly mean temperatures, deviations from the 10-year mean, averaged across stations, were calculated for the periods November–March, March–May and May–November each year, using:

$$\text{temperature deviation} = \frac{1}{p}\sum_{m}^{n}\left[\frac{1}{s}\sum_{s}(t_{ij} - \bar{t}_{ij})\right]$$

where t_{ij} = mean temperature at station j for month i; $\bar{t}_{ij}$ = 10-year mean temperature; s = number of stations with temperature data for month i; n = first month and m=last month of period; p = number of months in period.

These periods correspond to the pre-spawning (gonad development), spawning and growth seasons in bass, respectively (Kelley, 1988; Pawson and Pickett, in prep.). Sea temperature records for Bournemouth, which lies within the usual distribution range of the sampled juvenile bass population (Pawson *et al.*, 1987), are complete for the period 1979–1990, and were similarly used to investigate correlations with population parameters.

Year-class strength in exploited stock

Catch-at-age data for bass have been obtained annually from the commercial landings since 1983, using a catch-per-effort log-book scheme and biological sampling of

catches (Pickett, 1990), and these were examined in order to assess the recruitment of individual year classes to the fishery.

Results and discussion

Table 1 gives the strengths of bass year classes spawned in 1980–1989 relative to the 10-year mean. The high abundance of broods in 1981–1983 and virtual absence of fish spawned in 1985 were subsequently reflected in catch-at-age data obtained from the commercial fishery in Division VIId, Table 2.

Figure 2 shows the annual deviations of monthly mean sea surface temperatures from the 10-year (1980–1989) mean for the periods November–March, March–May, and May–November, as recorded from seven stations adjacent to the spawning and nursery areas of bass in the central English Channel. There is a clear pattern of warming through 1981–1983, cooling in 1984–1987, and a rapid warming through 1988 and 1989. Data for Bournemouth (Table 3) were used to show that there is a positive correlation between this temperature pattern and the indices of bass year-class strength (Fig. 3). Temperatures prior to and during the spawning season (March–May) were not so closely linked with the subsequent abundance of juvenile bass as those during the following summer and first winter of each year class.

Table 1. The relative abundance of the 1980–1989 year classes of bass in the Solent, and the mean length of the 1979–1988 year classes at the end of their third season of growth.

Year class	Relative abundance (survey mean c.p.u.e.)	Length at end of third year (cm)
1979	–	23.0
1980	0.44	24.5
1981	1.37	26.0
1982	1.51	26.0
1983	2.24	23.0
1984	0.36	23.8
1985	0.01	24.0
1986	0.04	–
1987	1.33	27.0
1988	1.84	26.0
1989	4.89	–

Table 3. Seasonal temperature deviations from the 1979–1990 11-year means in coastal waters at Bournemouth in the English Channel (°C per month).

	Nov-Mar	Mar-May	May-Nov
1979			
	0.52		
1980		0.07	−0.73
	0.11		
1981		0.37	−0.70
	−0.11		
1982		0.73	0.59
	0.53		
1983		−0.20	0.90
	0.17		
1984		−0.47	0.21
	−0.67		
1985		−1.00	−0.96
	−1.39		
1986		−2.27	−1.41
	−2.35		
1987		−2.30	−0.81
	−1.19		
1988		0.07	−0.19
	1.63		
1989		2.20	1.64
	2.85		
1990			1.46

The growth of juvenile bass in the Solent nursery area also appears to have been affected by this climatic pattern. The mean lengths reached by successive year classes of bass at the end of their third year (Table 1) are positively related to temperatures in the nursery area during the two preceding growth periods (length = 24.85 + 1.39 temperature deviation; r = 0.85). With the exception of the strong 1983 year class, which suffered two relatively cold growth seasons in 1984 and 1985, there is also evidence of a positive relationship between growth and year-class strength (r=0.87), supporting the suggestion that intra-specific competition and food availability are not usually limiting factors in the growth of juvenile bass, but that temperature is more important (Lancaster, 1992).

Table 2. Catches (thousands) of the 1980–1985 year classes of bass landed in ICES Division VIId in 1983–1989 (4-groups shown in bold).

	Year						
Year class	1983	1984	1985	1986	1987	1988	1989
1980	0.4	**140.6**	19.1	8.4	2.2	4.2	5.9
1981	0.2	14.0	**192.5**	42.7	16.8	12.2	5.6
1982	0	0	31.6	**61.9**	162.1	20.3	14.0
1983	0	0	0.6	16.1	**57.2**	128.1	41.8
1984	0	0	0	0	0.9	**18.8**	5.7
1985	0	0	0	0	0	0.3	**1.0**

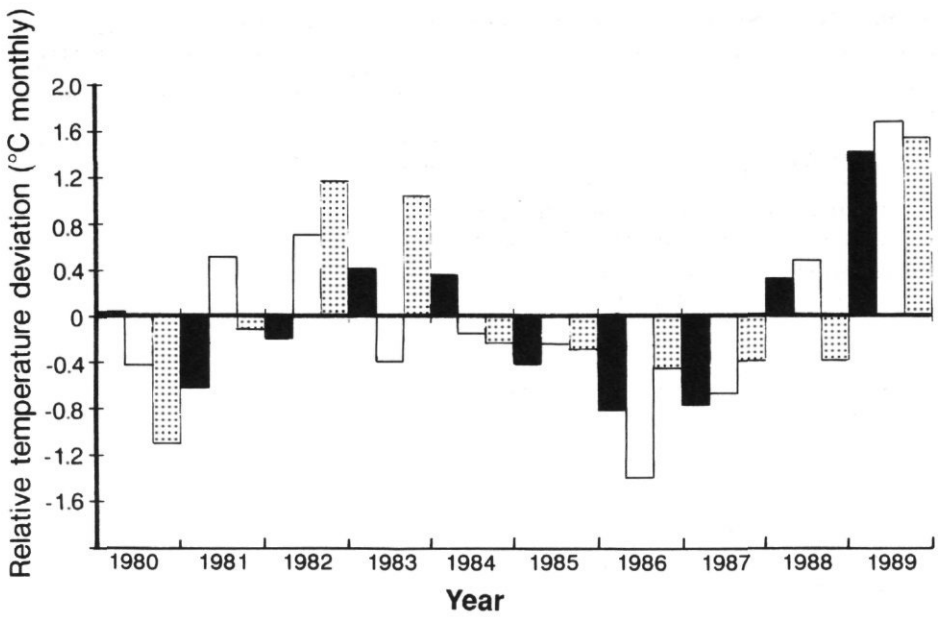

Figure 2. Seasonal temperature deviations from the 1980–1989 10-year mean in English coastal waters of the central English Channel: November to March (■); March to May (□); and May to November (⊞).

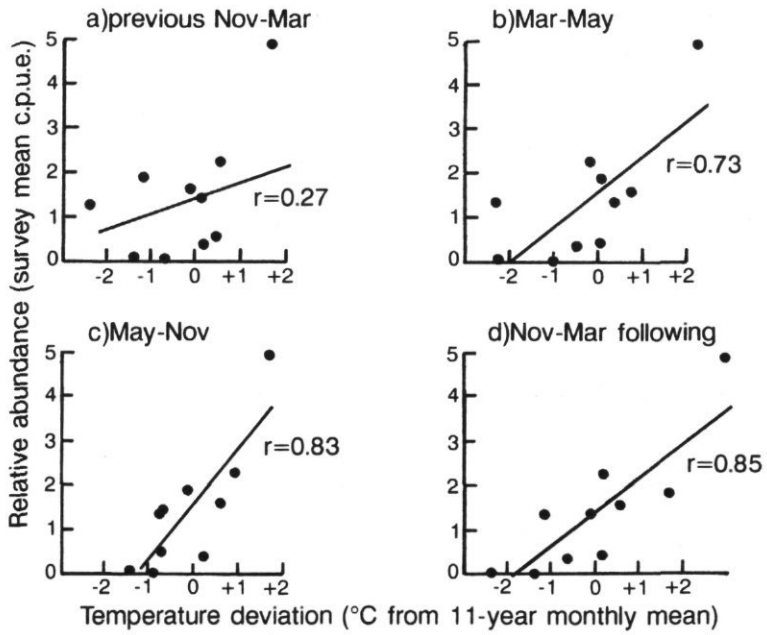

Figure 3. Relationships between relative year-class strength of bass and Bournemouth seasonal water temperature deviations (1979–1990). Lines fitted by linear regression.

Conclusions

These observations suggest that changes in the climate during the 1980s – as represented by temperatures of coastal sea water – have had a considerable influence on recruitment of bass to the fishery in the English Channel. It appears that any influence of temperature on the distribution of adult bass and their reproductive biology prior to and during spawning, and hence on the potential recruitment of larvae to nursery areas, has less effect on year-class strength than do growth opportunities during the first year and subsequent overwintering survival of zero groups.

Prior to 1983, bass in the Solent reached 36 cm (now the EC minimum landing size) as 5-groups, on average, but warming during 1981–1983 resulted in faster growth of the strong 1981 and 1982 year classes which reached 36 cm almost one year earlier. Growth of the strong 1983 year class in particular was depressed during the cooler period from 1984 to 1987, but warming in 1988–1990 should have enabled the above-average 1987–1989 year classes to recruit to the fishery as 4-groups. Thus, not only has climate influenced year-class strength of bass in nursery areas, it has also had additional effects on growth and the age of recruitment to the fishery, and appears to override any intra-year class density-dependent factors.

From a management viewpoint, earlier and stronger recruitment of year classes, which can arise from a period of relative warmth, will improve yields in the fishery and boost the spawning stock, provided that the incoming recruits are not targeted too heavily. Technical measures aimed at protecting bass under 36–40 cm (Pawson and Pickett, 1988) have recently been implemented around the UK and in Eire for this purpose.

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A re-examination of the relationship between cod recruitment and *Calanus finmarchicus* in the North Sea

K. Brander

Brander, K. 1992. A re-examination of the relationship between cod recruitment and *Calanus finmarchicus* in the North Sea. – ICES mar. Sci. Symp., 195: 393–401.

Recruitment of cod and other gadoid species was high during the late 1960s and the 1970s and declined throughout the 1980s. This so-called "gadoid outburst" has been ascribed by Cushing (1984) to changes in the abundance and timing of peak occurrence of *Calanus finmarchicus* in the northeastern North Sea. Continuous Plankton Recorder (CPR) data for that area show no marked trends in abundance and timing during the period 1958–1990, and the multiple regression proposed by Cushing does not model observed variability in cod year classes. Since cod spawn over much of the North Sea and their larvae feed on a range of sizes and species of copepod, a more complete model of the relationship between their abundance and the survival of cod larvae is needed. The CPR may provide some of the extensive population data on copepods which would be required for such a model.

K. Brander: Ministry of Agriculture, Fisheries and Food, Fisheries Laboratory, Lowestoft, Suffolk NR33 0HT, UK.

Introduction

The total catch of four large gadoid species (cod, haddock, saithe, and whiting) in the North Sea increased from 200–300 Ktonnes during the period 1948–1963 to 1500 Ktonnes in 1969 and has since declined fairly steadily to about 400 Ktonnes in 1990 (Fig. 1). The increase in catch was due to increased biomass, in particular the outstanding 1967 haddock year class and also higher levels of cod recruitment than in previous decades. The subsequent slow decline (although catches

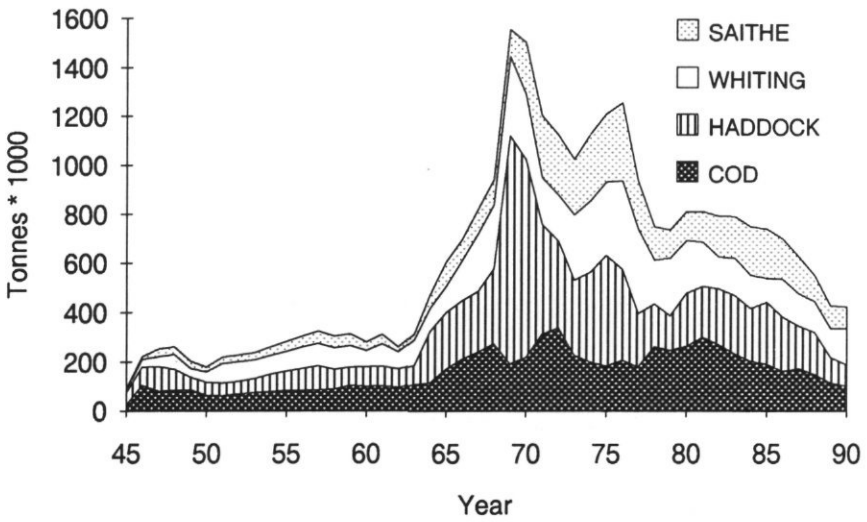

Figure 1. Total international catch of large gadoid species (cod, haddock, whiting, and saithe) fro... the North Sea.

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remain higher than during the earlier period) is due to reduced biomass and increased fishing mortality (Daan *et al.*, 1990). Attempts have been made to look for general explanations for the so-called "gadoid outburst" and to associate it with long-term changes in the physical and biological environment. Cushing (1984) suggested that: "The upsurge of cod-like fishes from 1962 onwards is then attributed to the delay in *Calanus* production in the northeastern North Sea by about a month." He reported a correlation between the spring timing and abundance of the copepod *Calanus finmarchicus* in the NE North Sea over the period 1954–1977 and the subsequent recruitment of cod.

The annual catch of cod in the North Sea rose from around 100 Ktonnes during the period 1948–1962 to over 300 Ktonnes in 1971 and 1972 due to the large 1969 and 1970 year classes and declined throughout the 1980s to just over 100 Ktonnes in 1989 (Fig. 2). The average recruitment (number of one-year-old fish estimated by VPA; Anon, 1991) was 20–30% higher during the 1970s than during the 1960s or 1980s, but the interdecadal differences are not significant because of the annual variability (Fig. 3). Trends in the catch of cod depend largely on the level of recruitment (Fig. 4), so we need to understand the causes of variability in early life, probably during the pelagic phase, in order to account for the observed changes in catch. Surveys of juvenile cod shortly after they settle from the plankton to the seabed give an indication of recruitment, but the degree to which year-class strength is determined by feeding and survival during the first four months is not known. Recent studies (Ellertsen *et al.*, 1989) support Cushing's hypothesis that the growth and survival of cod larvae depend on both the timing and abundance of *Calanus finmarchicus*. The correlation which Cushing reported can be investigated in terms of component processes rather than being purely phenomenological.

The aim of this paper is to review the evidence relating cod recruitment to *Calanus* abundance in the North Sea over the period 1960–1989 and to discuss the component processes. This leads inevitably to spatial and temporal disaggregation, because the distribution of cod larvae and their prey in space and time varies from year to year. The information from the CPR surveys is evaluated to see whether it may provide the spatial and temporal coverage and resolution of copepod populations required to model their dynamics.

Materials, methods and results

The CPR data used by Cushing (1984) were the numerical abundance of *Calanus finmarchicus* plus *helgolandicus* for the northeastern North Sea (area B1, 58–63°N and east of 3°E) for the months April–June for the period up to 1978. The original counts were transformed ($\log(1 + x)$) and backtransformed using an empirical relationship described by Colebrook (1975). The time series has been updated (Table 1), but years with missing data (i.e. not all months sampled) have been excluded. The average abundance of *Calanus helgolandicus* is less than 5% of *Calanus finmarchicus* in the northeastern North Sea area at this time of year.

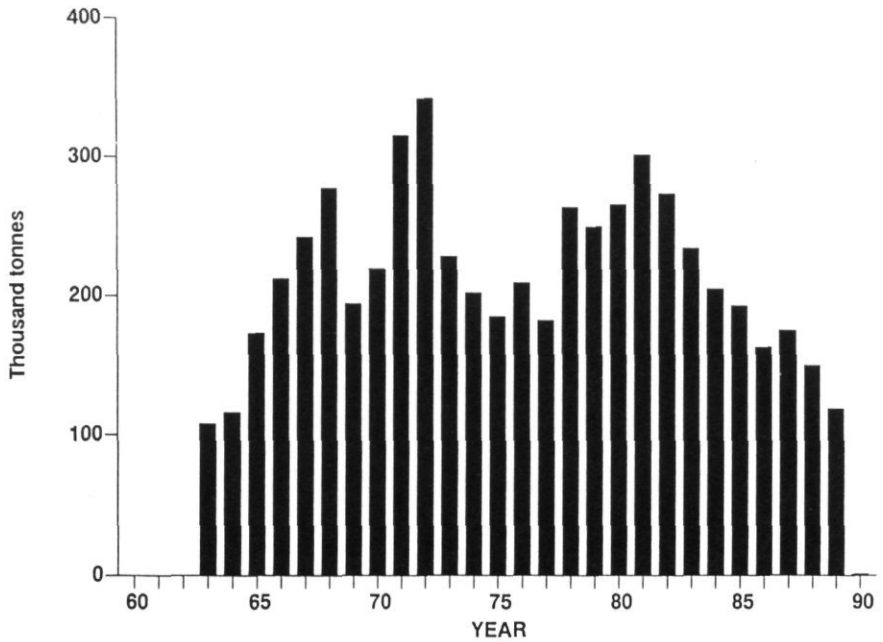

Figure 2. Total international catch of cod from the North Sea (Anon., 1991).

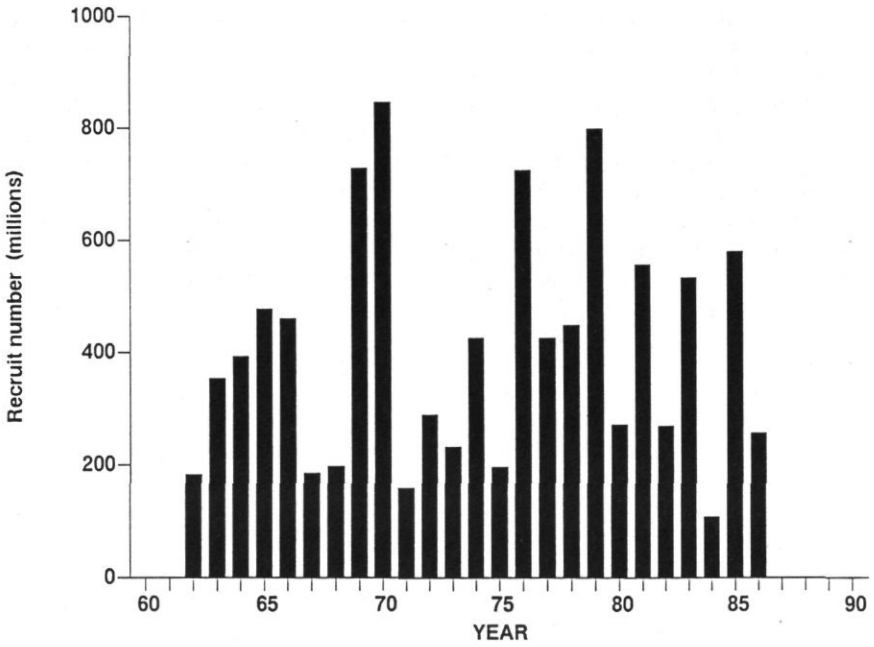

Figure 3. Year classes of cod (number of one-year-old fish) estimated by VPA (Anon., 1991).

In order to provide a more complete picture of distributional and secular changes in *Calanus* spp. over the whole of the North Sea, CPR data for each ICES rectangle (1° longitude × 1/2° latitude) for every month between 1958 and 1990 were included in the analysis. Details of the CPR sampling are given by the Oceanographic Laboratory, Edinburgh (1973). The number of monthly samples in each rectangle over the 30-year period (1960–1989) is given in Figure 5 and the annual mean number of *Calanus* (copepodite stages V + VI m^{-3}) in Figure 6. Sampling was most frequent along the regular routes, but some of these have changed over time. In order to simplify analysis and interpretation and to ensure adequate numbers of samples in each area and time period, the North Sea was divided into four quadrants (see Fig. 5). The mean number of *Calanus* in each

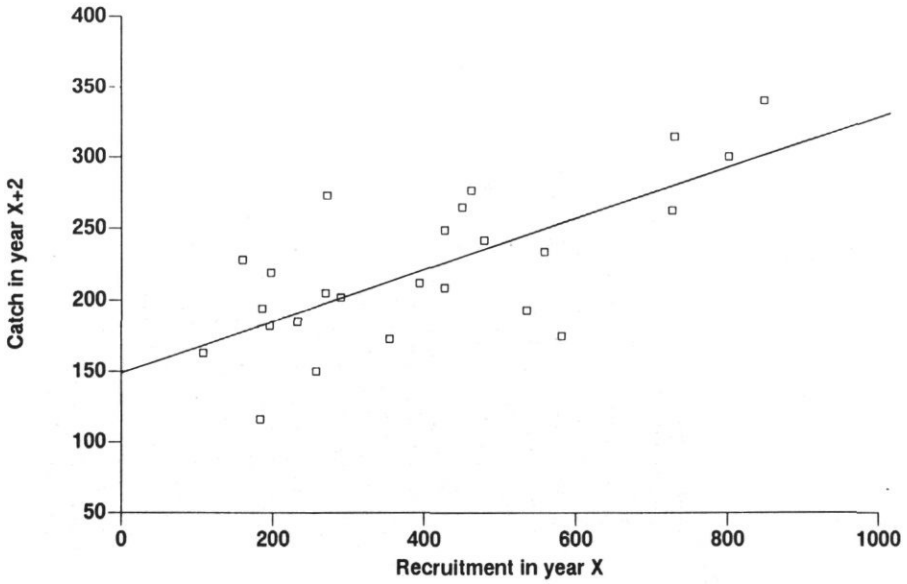

Figure 4. Catch of cod in the North Sea as a function of the number of one-year-old recruits (estimated by VPA), lagged by two years ($r^2 = 0.50$, 25 d.f.).

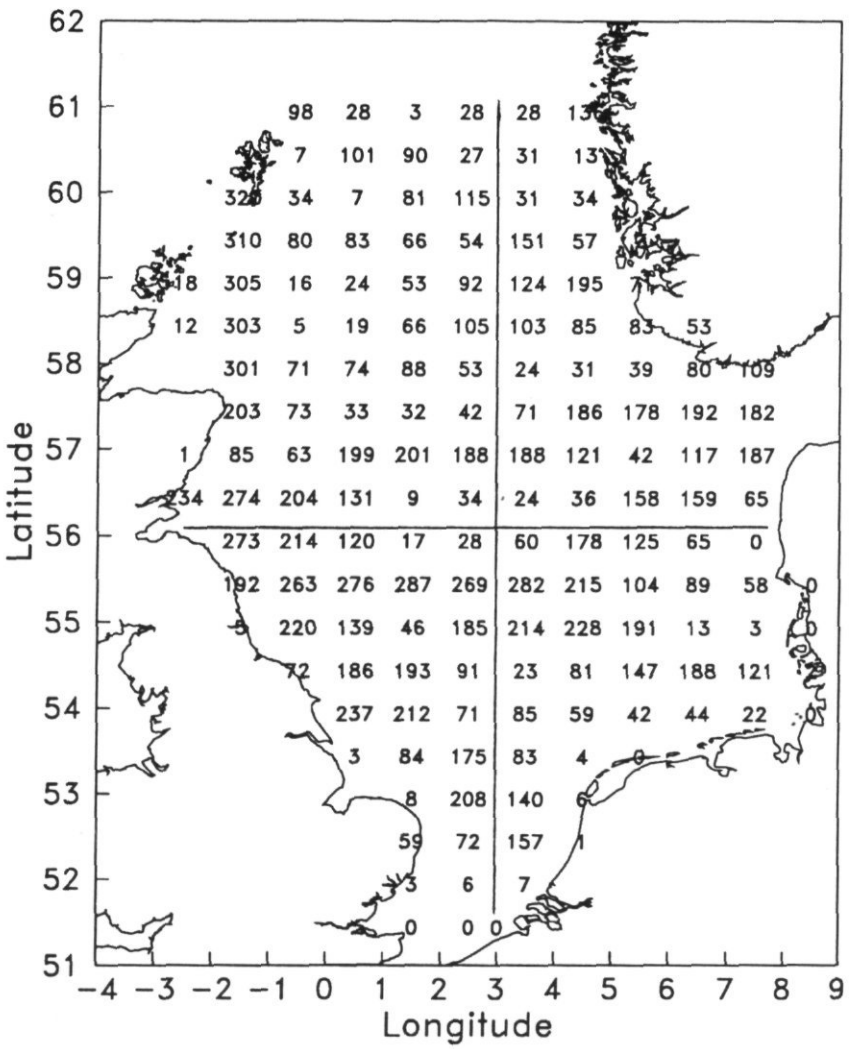

Figure 5. Number of months in which each rectangle was sampled during the period 1960–1989. The four quadrants into which the whole area has been divided for analysis are marked.

of these four areas in each month (Fig. 7) gives an indication of the variability between years, between areas, and seasonally. The monthly abundances for the four areas are all significantly positively correlated, but the closeness of the relationship varies considerably and it must be noted that the seasonal signal is included in the analysis.

Coefficient of determination ($r^2 > 358$ deg. of freedom) of monthly rectangle values for four quadrants in the North Sea.

	NE	NW	SW	SE
NE AREA		0.22	0.07	0.02
NW AREA			0.23	0.09
SW AREA				0.31

The mean annual abundance of *Calanus* fluctuated without evident trend over the 30-year period (Fig. 8). The mean abundance during spring (February to May) may have declined slightly until the end of the 1970s and then increased again (Fig. 9). The timing of *Calanus* abundance during the first six months of the year was calculated by weighting the date (in months) by the abundance in each month and dividing by the total abundance. There was no evident trend in timing in the two northern areas, but in the southern area the timing advanced by about two weeks over the 30-year period (Fig. 10).

Cushing (1984) applied his multiple regression model to data for the years 1954 to 1977 and found a correlation coefficient of 0.559 for 21 d.f. ($p < 0.05$). The same model was applied to the *Calanus* data for area B1 and

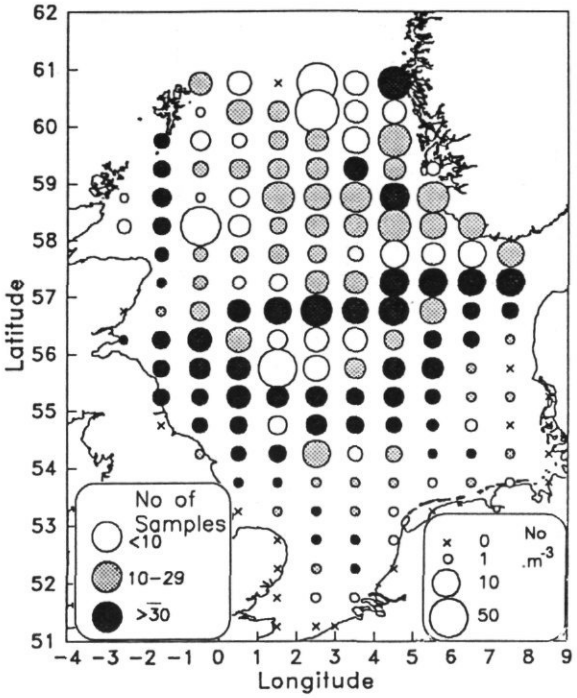

Figure 6. Mean monthly catch (number m^{-3}) of *Calanus finmarchicus* stages V + VI in each rectangle averaged over the period 1960–1989. The shading of the symbols shows the level of sampling in each rectangle.

Table 1. Delay in *Calanus* (month after March in which peak abundance occurs); *Calanus* abundance (mean no. m^{-3} for area B1 in April–June); cod recruit number (millions of one-year-olds from VPA); Atlantic inflow (% of North Sea north of 57°N covered by Atlantic water >35.1 psu at 50–200 m – from Svendsen and Magnusson (1992)).

Year	Delay	*Calanus* m^{-3}	Cod recr't	Inflow
1962	2	32.25	183	
1963	2	42.09	354	
1964	3	49.96	393	
1965	3	53.23	478	
1966	2	34.71	461	
1967	2	27.80	185	
1968	3	21.74	197	76
1969	3	59.12	729	37
1970	3	63.57	847	45
1971	2	28.23	159	56
1972			289	55
1973	3	46.44	232	48
1974			426	44
1975			196	36
1976	2	46.13	726	47
1977		27.54	426	25
1978	2	30.64	449	10
1979	2	20.58	800	29
1980	3	22.13	271	23
1981	2	4.57	557	35
1982	3	6.72	269	38
1983	3	19.59	534	38
1984	2	49.26	108	34
1985	1	34.15	581	NA
1986	2	18.63	257	44

the most recent estimates of cod recruitment (Anon., 1991) for the year classes 1962–1986. The year classes from 1954 to 1961 were not included because they are not part of the VPA time series.

Cod recruitment
= Fn (Delay in *Calanus* timing, *Calanus* abundance)

Analysis of variance for the multiple regression of cod recruitment on *Calanus* timing and abundance. Data used are given in Table 1.

Source	Sum of squares	Degrees of freedom	Mean square	F
Delay	99 366	1	113	0.0
Calanus	11 369	1	83 180	1
Del*Cal	118 761	1	126 901	2.7
Residual	849 742	17	47 511	
Total	1 079 237	20		

Neither of the main covariates (abundance and timing of *Calanus*) nor their interaction accounted for a significant part of the variance in recruitment. This is also the case if the tests are carried out stepwise. In spite of the failure of this simple model, it is probably premature to abandon investigation of the relationship between cod recruitment and copepod abundance. The model is incomplete in several respects, since it treats the whole North Sea stock as one unit which is dependent on the abundance of one species of copepod in one area. The relationship between cod recruitment and *Calanus* abundance, if it exists, is not strong enough to emerge from this simple analysis.

Cushing (1984) also suggested that variability in Atlantic inflow into the North Sea might be responsible for observed fluctuations in *Calanus finmarchicus*, since the distribution of this species is principally oceanic. Svendsen and Magnusson (this symposium) provide an index of the influx of Atlantic water into the northern North Sea for the period 1968–1990. This is the percentage of the North Sea north of 57°N over which the salinity at a depth of 50–200 m is greater than 35.1 psu (see Table 1). The measurement is made in June/July, but they regard it as a good index of the integrated flux during the previous months. The correlation between this index and the abundance of *Calanus* in area B1 in April–June is not significant.

A more complete model of larval population processes should include more detail of the distribution of spawning and consequent larval distribution of cod. This distribution has been well described for the southern North Sea (Daan, 1978), but information for the rest of the North Sea is scant. From the ICES pelagic 0 group

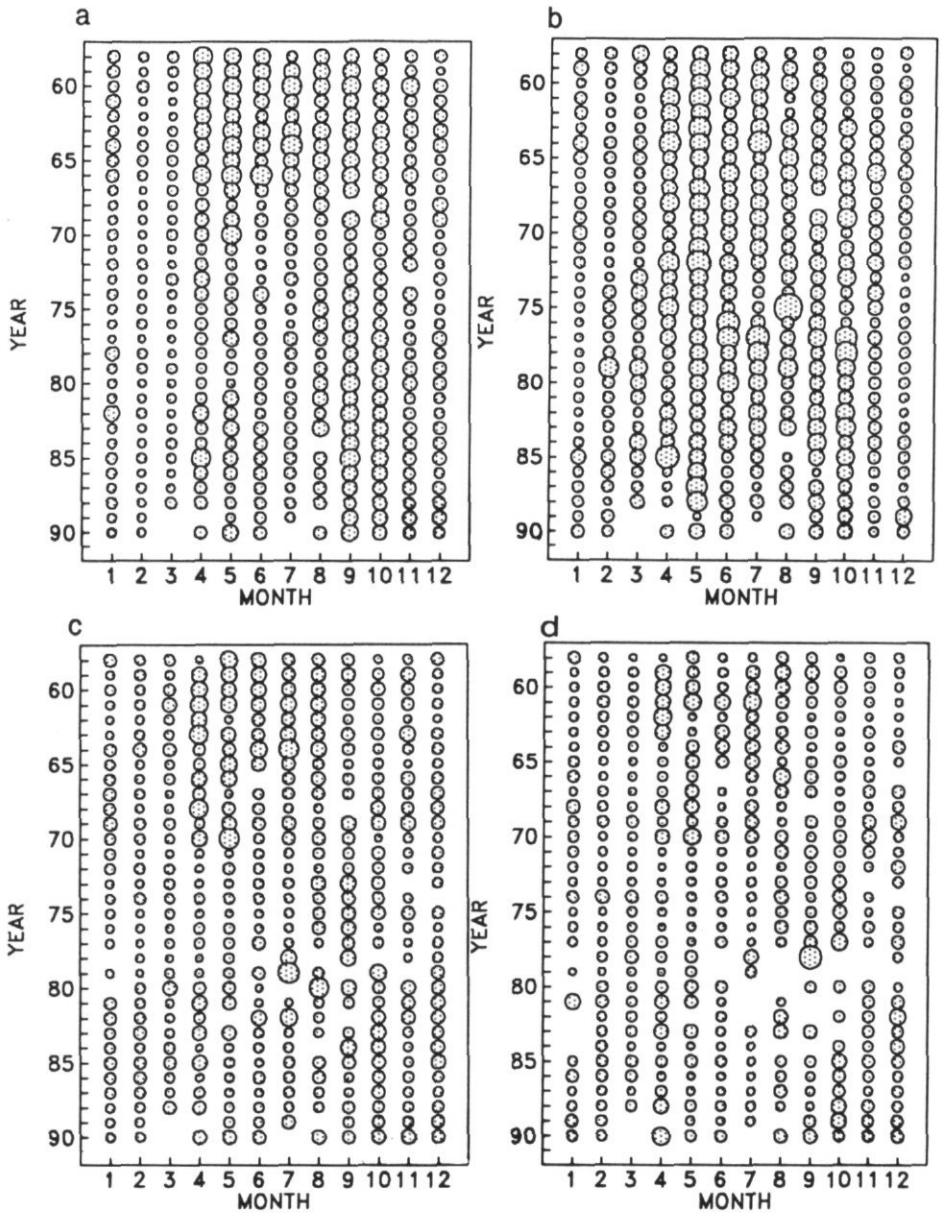

Figure 7. Mean number m^{-3} of *Calanus* in each quadrant ((a) NW, (b) NE, (c) SW, (d) SE) of the North Sea in each month for the period 1960–1989.

surveys over the period 1974–1983 it appears that the main area of occurrence of cod larvae is in a wide arc, west and south of the Norwegian Trench, extending north of 61°N (Fig. 11). This differs considerably from the distribution of cod larvae given by the CPR for 1948–1956 (Fig. 12), when the largest numbers occurred in the German Bight and NW area of the North Sea, but not north of 59°N. Coombs (1980) noted that cod larvae were absent from CPR samples in the German Bight between 1958 and 1972 and there may have been a northerly shift in the distribution of spawning since 1950.

Discussion

The rationale behind the multiple regression model relating cod recruitment to *Calanus* abundance and timing is the match-mismatch hypothesis (Cushing, 1990). There are a number of difficulties in applying it to a stock such as North Sea cod, in which spawning occurs over a wide area and over several months. The hypothesis postulates a relationship between production of suitable food and survival of larvae. Although cod feed principally on copepods throughout the larval stage

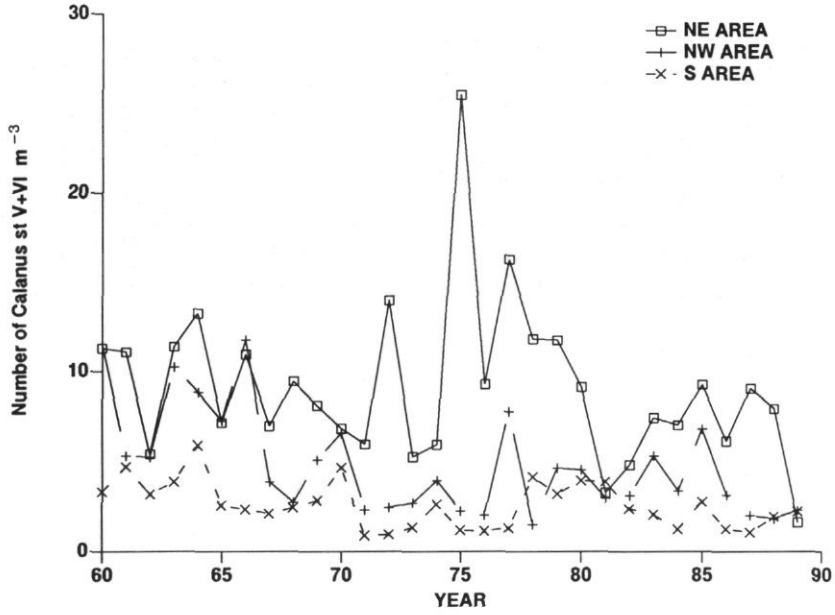

Figure 8. Mean annual abundance of *Calanus* stages V + VI.

(Ellertsen *et al.*, 1989), the size composition of their prey increases and they feed on different species in different areas.

For most of their larval life cod feed on stages of *Calanus* which are smaller than those whose distribution is described here (the adult stages V + VI). The CPR, with mesh size of 285 × 315 microns, does not sample small stages of *Calanus* adequately (Nichols and Thompson, 1991). However, the production of eggs and nauplii of *Calanus* in spring is a function of the number of adult copepods, as well as of temperature and primary production; therefore adult abundance may be used to estimate food available for fish larvae.

A more detailed model relating cod recruitment to copepod dynamics for the North Sea should include four main components:

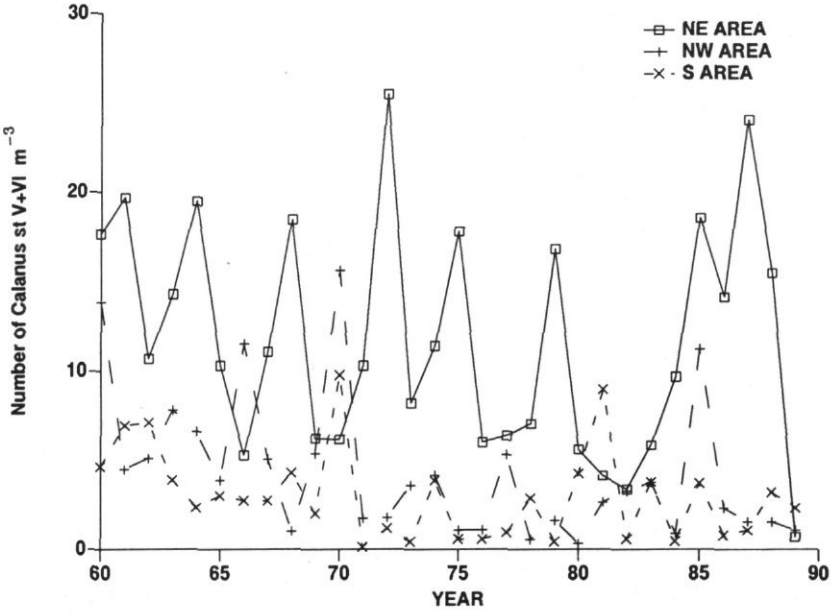

Figure 9. Mean spring (Feb–May) abundance of *Calanus* stages V + VI.

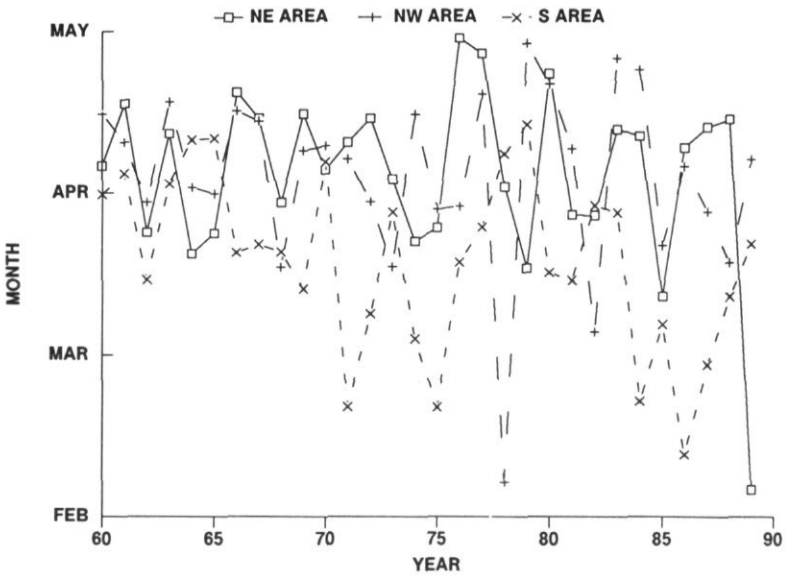

Figure 10. Mean timing of *Calanus* occurrence during the first six months of the year.

1. Distribution of larvae in space and time.
2. Distribution of major copepod food species.
3. A model of copepod egg and nauplius production as a function of temperature, advection, adult copepod abundance, and phytoplankton production.
4. A model of the relationship between copepod abundance and larval survival.

The first of these would be extremely costly in survey time for an area as large as the North Sea. Dispersion and advection of larvae could be modelled after the initial distribution had been determined by survey, but resulting population distributions also require assump-

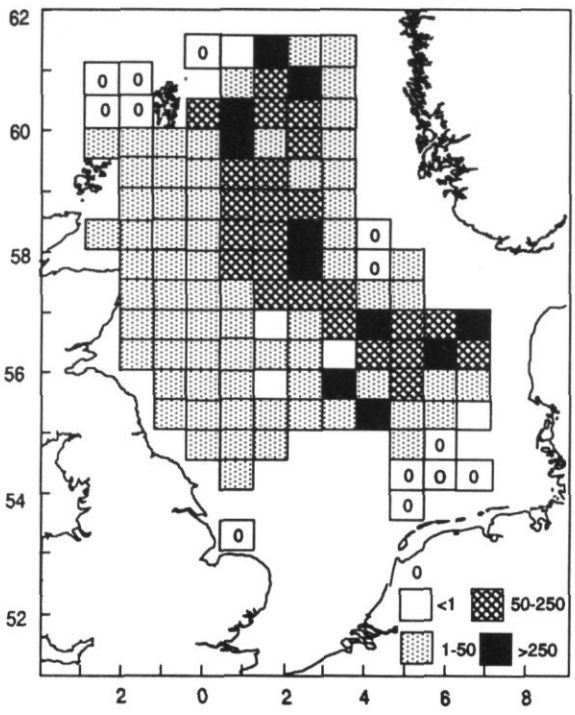

Figure 11. Mean distribution of pelagic zero-group cod in June/July for the period 1974–1983.

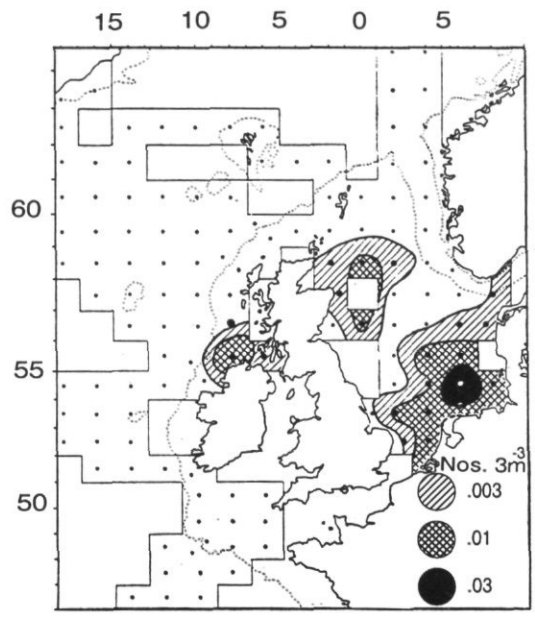

Figure 12. Mean distribution of cod larvae from the CPR for the period 1948–1956 (from Henderson, 1961). Cod larvae occur in all months from February to August, with greatest numbers in April.

tions about mortality rates. Modelled distributions may be validated using CPR data, although the volumes sampled by the CPR are small in relation to concentrations of fish larvae.

For copepods the CPR sampling is clearly of great value in showing distribution patterns. Its usefulness for cohort-based population analysis is more limited, because it does not sample all sizes of copepods.

A geographic array of one-dimensional models of copepod production as a function of temperature, primary production, and adult population could be constructed for the North Sea in order to resolve time and space scales which affect the overall population dynamics of fish larvae. If transport of overwintering adult populations is a major cause of year-to-year variability then it may be necessary to include horizontal processes. The CPR provides geographic coverage of adult population levels; however, it is not obvious that the data allow rates of horizontal transport to be estimated. Indicator species provide qualitative information on major circulation changes. Cushing's multiple regression model resolves the time scale by including the timing of peak abundance of *Calanus*, but it ignores spatial heterogeneity.

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Natural changes in pelagic fish stocks of the North Sea in the 1980s

A. Corten and G. van de Kamp

Corten, A., and Kamp, G. van de. 1992. Natural changes in pelagic fish stocks of the North Sea in the 1980s. ICES mar. Sci. Symp., 195: 402–417.

One of the conspicuous natural changes in North Sea fish stocks in the 1980s was a shift in summer distribution of the adult herring. During the 1970s and earlier decades, the main feeding grounds in June/July were the waters off the east coast of Scotland south of 57°N and west of 2°E. During the 1980s, very few herring were found in this area during the feeding season. The distribution of the adults had shifted north and northeast, to the edge of the continental shelf and to the Norwegian Trench. Herring recruitment steadily increased during the early 1980s, and was above the long-term average from 1984 to 1986. During the years of strong recruitment, a relatively high proportion of North Sea juveniles grew up in the Skagerrak/Kattegat. The change in adult herring distribution corresponded in time and area with earlier documented changes in western mackerel. In both species, the summer distribution shifted towards the northeastern North Sea. Considering the similarity in time and area, it is assumed that the changes in herring and mackerel had the same hydrographic origin. The changes in both species occurred at the edge of the continental shelf, in the region of the shelf edge current. It is therefore suggested that changes in the shelf edge current might have been responsible for the observed changes.

A. Corten and G. van de Kamp: Netherlands Institute for Fisheries Research, PO Box 68, 1970 AB IJmuiden, The Netherlands.

Introduction

The interest among fisheries biologists in natural changes in North Sea fish stocks seems to have a long-term periodicity in itself. The discussion about possible natural changes in North Sea pelagic stocks was very lively at the end of the last century (Ljungman, 1882), and also in the period after the Second World War (Kreft, 1963; Zijlstra, 1963). However, the interest in this subject decreased in later years (with the exception of the ICES Aarhus symposium in 1975) and in the period 1970–1985; few ICES assessment working groups dealing with North Sea stocks considered the possibility of changes being brought about by variations in the environment.

After a period of minimal interest, the subject of natural change started to attract attention again in the mid-1980s. Some fish stocks in the North Sea and to the west of the British Isles varied in a way that could not be explained by fishery effects or species interaction, and scientists were forced to consider the possibility of natural changes again. A major encouragement to do so was the publication of Cushing's (1982) book on climate and fisheries, which clearly illustrated the world-wide occurrence of natural changes in fish stocks.

In a paper on changes in herring recruitment, Corten (1984) showed that the prolonged period of recruitment failure in 1974–1978, and the steady improvement in later years, could not be ascribed merely to changes in parent stock biomass. Using data on distribution of five-month-old larvae, he demonstrated that during the period of recruitment failure the larvae had not reached their normal nursery area in the eastern North Sea. On the basis of this observation, he speculated that the residual circulation in the North Sea had been changed during a period of several years.

In 1985 the ICES Mackerel Working Group presented a detailed description of changes in mackerel distribution west of the British Isles and in the northern North Sea (Anon., 1985). They described a shift of mackerel overwintering areas from the area southwest of England to the waters west of Ireland, and then to the area north of Scotland. At the same time the summer migration of western mackerel extended further into the North Sea, and the distribution area of juveniles was extended northward. No conclusions were drawn as to the cause of the observed changes, but it was obvious that the shifts in distribution area were not related to the effects of fishing, or to changes in stock size or age composition.

Walsh and Martin (1986) provided further details on

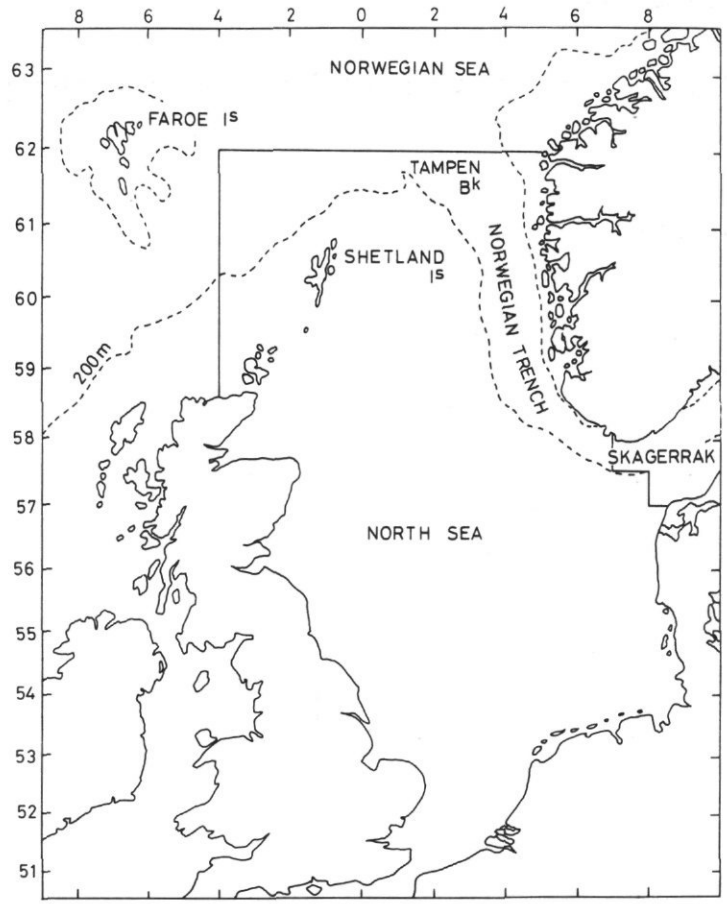

Figure 1. Study area with geographical names mentioned in text.

the changes in mackerel distribution, particularly during the autumn fishery west of Scotland. They concluded that the northern displacement of the western mackerel stock was probably related to long-term hydrographic changes, involving an increased flow of Atlantic water along the shelf edge of western Europe.

The Sprat Biology Workshop in 1986 (Anon., 1988a) considered the dramatic changes in sprat abundance and distribution in the North Sea during the 1970s and early 1980s. The sprat stock in the North Sea expanded strongly in the early 1970s, and the centre of gravity of the population shifted from the southeastern North Sea to the western North Sea. By the end of the decade, the stock decreased just as suddenly as it had developed in the earlier years, and the entire sprat fishery in the western North Sea collapsed. During the years of the decline (1978–1982), there was a southward shift in the distribution of the sprat stock. The Sprat Biology Workshop was unable to explain these changes in stock size and distribution on the basis of fishery effects, concluding that the changes were probably linked to a long-term environmental change.

Corten (1990) reviewed natural changes in herring, sprat, mackerel, and plankton during the 1970s and 1980s, and tried to link all these changes by a theory that postulated a reduction of Atlantic inflow into the northwestern North Sea in the 1970s and an increased inflow in subsequent years.

In the present paper, attention is focused again on changes in the herring stock. Whereas earlier papers mainly dealt with changes in larval distribution and recruitment, this paper will consider primarily the distribution of adult herring. During the 1980s, fishermen frequently reported an abnormal distribution of herring catches. At first, little attention was paid to these reports. It was assumed that changes in herring distribution were due either to random fluctuations from year to year, or to changes in herring stock size and age composition. However, the reports on northerly catch distributions continued for a number of years, and were later confirmed by the results of international acoustic surveys. At this stage it became clear that a closer look at the distribution of the adult stock was necessary.

This paper compares changes in herring distribution

with the earlier described changes in mackerel. It investigates whether a common pattern exists between the two stocks, and what mechanism could be responsible for the observed changes. The area in which the fish stocks were studied is shown in Figure 1.

Changes in North Sea herring

Distribution of adult herring

After the period of stock depletion and closed fishing in the late 1970s, the North Sea herring stock showed a remarkable recovery during the 1980s. The spawning stock increased from only 50 000 t in 1977 to an estimated 1.4 million tonnes in 1990 (Fig. 2). This recovery was the result of improved management and also of a series of strong year classes.

In the years following the re-opening of the fishery in 1983, Dutch fishermen frequently commented upon the unusual distribution of the herring. According to their reports, herring were absent from the traditional fishing grounds in the western North Sea during the summer. Instead, fishermen had to go very far north in order to find the herring. In these northern areas, herring were hard to catch with a pelagic trawl because they were swimming close to the surface.

In order to investigate possible changes in the distribution of North Sea herring, the monthly distribution of commercial catches in June and July has been considered. It was especially in this part of the fishing season that changes in distribution were being reported by the fishermen. At this time of the year the herring is still feeding, and its distribution is probably governed by the (variable) distribution of food organisms.

The distribution of herring catches in June and July has been considered over the period 1955–1990. Most attention was given to the data for the latest decade, as this was the period in which changes were expected to have occurred. Data on monthly catch distributions were derived from the sources indicated in Table 1.

No catch distributions were available for the years when the fishery was closed (1977–1982), and for the first year after the re-opening of the fishery. International statistics on catch distribution were presented by the ICES Herring Assessment Working Group only from 1985 onwards.

Table 1. Sources of monthly catch data.

Period	Type of data	Source
1955–1960	International catches	ICES *Annales Biologiques* 1955–1960
1970–1977	Dutch catches	Dutch national statistics
1985–1990	International catches	ICES Herring Assessment Working Group Reports (Anon., 1986, 1987, 1988b, 1989a, 1990a, 1991a)
1984–1990	Dutch catches	Dutch national statistics

For the 1970s, no international data were available, and the only source of information was the catch distribution of the Dutch fleet. The same type of data have also been included for the 1980s (in addition to international statistics), to make a direct comparison between catches taken in the 1970s and 1980s by the same fleet.

Prior to 1980 the monthly catch distributions have been averaged over the periods 1955–1960 and 1970–1977. Data for more recent years have been averaged over shorter intervals (1984–1987 and 1988–1990) in order to study possible trends over time in more detail. The resulting distributions for each month and period are presented in Figure 3.

When comparing the distribution of herring catches in the various periods it becomes obvious that drastic changes have occurred in recent years. Before 1980 the situation seems to have been relatively stable. The distribution of international catches in 1955–1960 is little different from the distribution of Dutch catches in 1970–1977. In the latter period, catches had shifted slightly to the west and south in comparison to the data for 1955–1960. This difference, however, may be due to a difference in average fishing position between the Dutch fleet and the combined international fleet. In both sets of

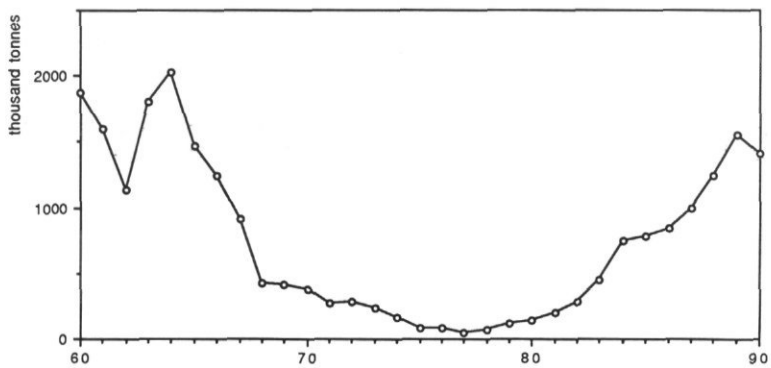

Figure 2. Spawning stock size of North Sea herring 1960–1990. Data from Anon. (1991a) and earlier working group reports.

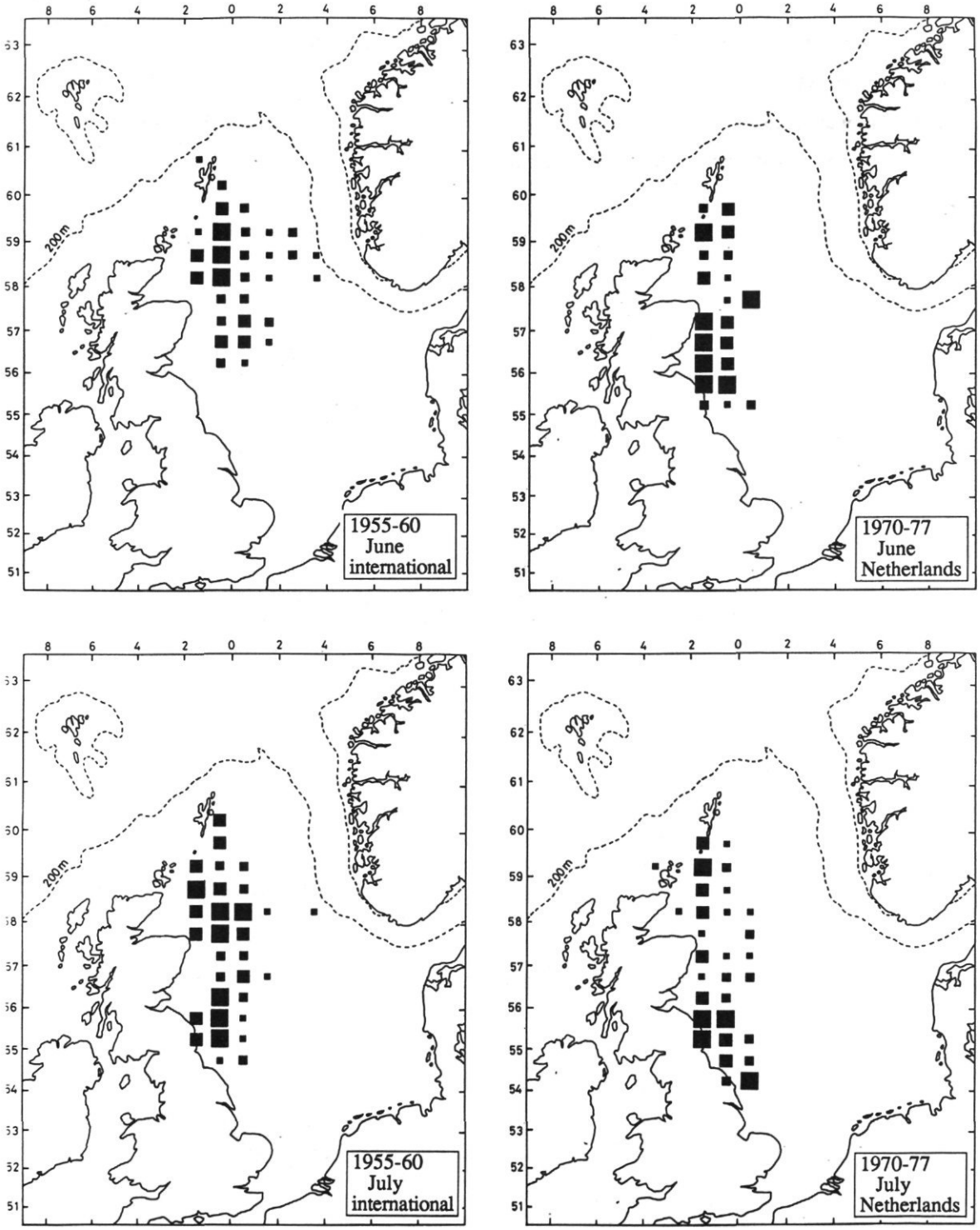

Figure 3. Distribution of herring catches in June and July for different periods and different fleets. Black squares represent, in order of increasing size, monthly catches of 250–500, 501–1000, 1001–2000, and >2000 tonnes for the international fleet, and 100–250, 251–500, 501–1000, and >1000 for the Dutch fleet. Catches west of 4°W are not included.

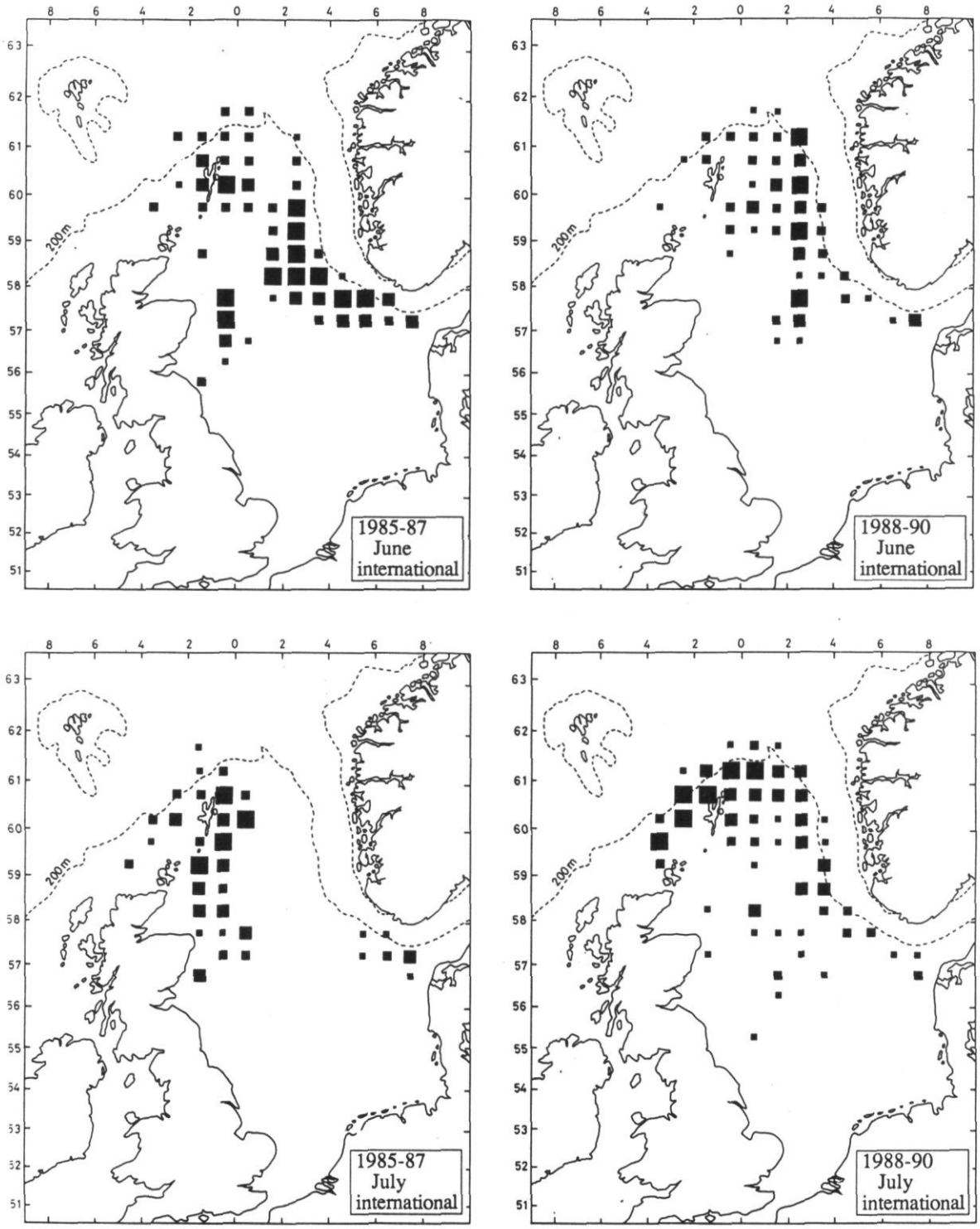

Figure 3 (*cont.*)

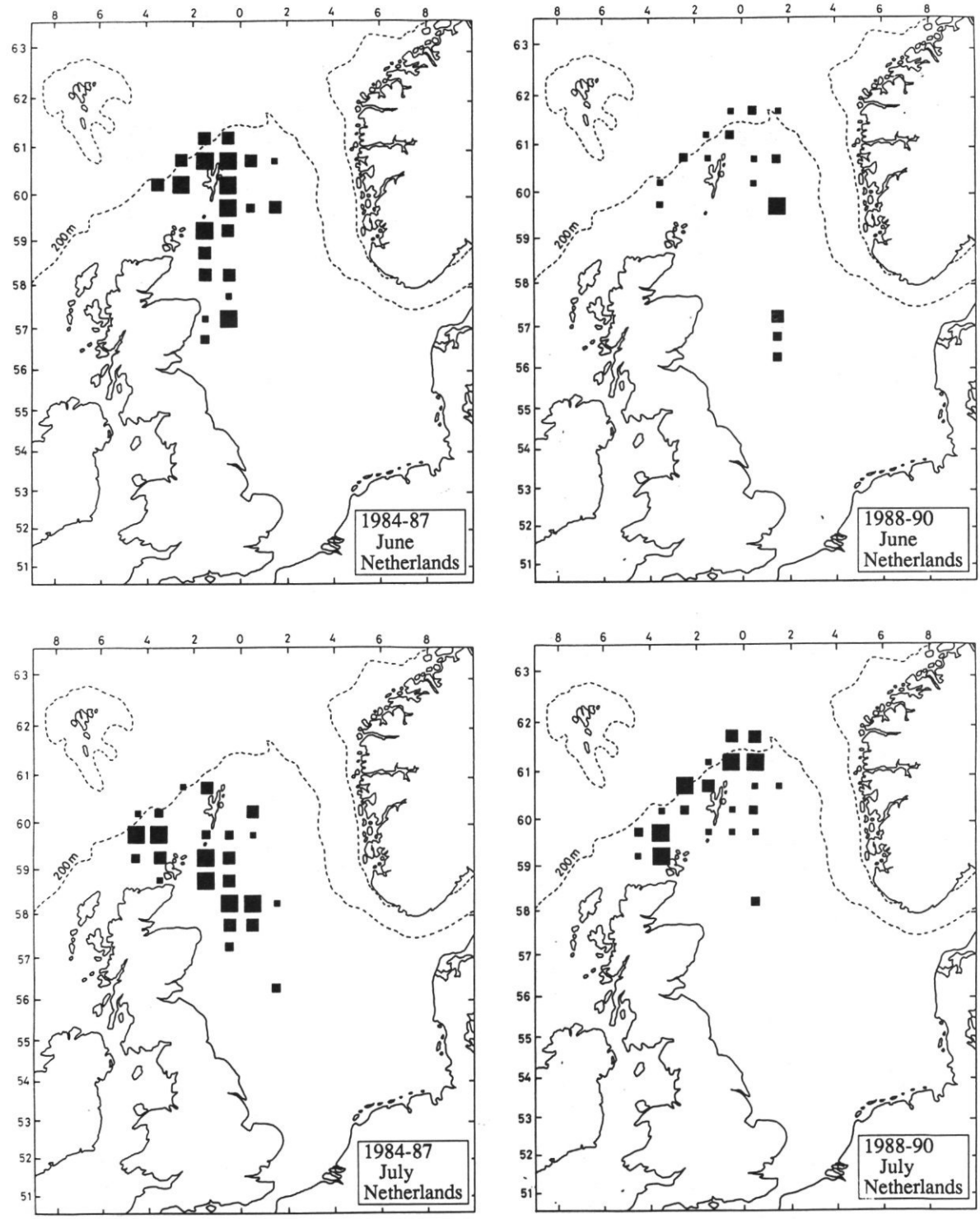

Figure 3 (*cont.*)

data, the catches in summer are taken in the western North Sea south of Shetland, and there is a slight southward shift in July, indicating the start of the migration towards the spawning grounds in the central North Sea.

A completely different picture is seen for the years after 1980. The international catches show a distinct northeasterly distribution in June and a northerly distribution in July. This pattern is most pronounced during the last years of the decade (1988–1990). In particular the distribution in June along the western edge of the Norwegian Trench is remarkable. Catches in this month on the traditional fishing grounds in the western North Sea are small and virtually disappear towards the end of the decade.

During July there is a shift of catches to the west and northwest, but not to the south as in earlier decades. There is no sign yet of the beginning of the migration towards the spawning grounds in the central North Sea. The distribution for July 1988–1990 is most unusual in that it shows all catches to be taken in the northern North Sea, in the region of the shelf edge, whereas there were virtually no catches on the traditional fishing grounds further south.

The distribution of the catches in the mid-1980s may have been influenced to some degree by a fisheries conflict between Norway and the European Community, which forced Norwegian fishermen to fish herring mainly in the Norwegian sector. The fisheries dispute, however, was resolved in 1986, and cannot have affected the distribution of catches in the later years of the decade.

The fact that the changes in fishing position between the 1970s and 1980s were due to a shift in herring distribution and not to national restrictions on fishing fleets is also demonstrated by a comparison of Dutch catches in both decades. The trawl fisheries of the Netherlands were traditionally conducted in the western North Sea during the summer. In the mid-1980s the fishery in June showed a pronounced northeasterly shift, and towards the end of the decade the catches in this month declined sharply. The only significant catches were taken along 1–2°E. As is shown from the distribution of international catches, most of the adult herring at this time were in the eastern North Sea, and Dutch fishermen for a variety of reasons did not search for herring in this sector. For the month of July, the Dutch catches in the 1980s show the same pattern as the international catches, with a very northwestern distribution in the period 1984–1987, and an extremely northern distribution along the shelf edge in 1988–1990.

To summarize the review of catch distributions in the last four decades, it may be concluded that the distribution pattern was relatively constant from 1955 to the beginning of the fisheries closure in 1977. After the reopening of the fishery in 1983, the herring appeared to be distributed very northerly and northeasterly. This abnormal distribution pattern became most pronounced in the late 1980s. A comparison of Dutch statistics for the 1970s and 1980s shows that the shift in distribution of the fishery also occurred within a single fleet, and that the abnormal distributions in the 1980s were not the result of national restrictions on the fleets of individual nations.

The very northern distribution in summer during the late 1980s is confirmed by the results of the ICES acoustic surveys in the North Sea. Reports from these surveys (Anon., 1989b; Kirkegaard *et al.*, 1990; Anon., 1991b) show that in July 1989–1990 most adult herring were found along the northern edge of the North Sea, or even beyond it. Very few adult herring were recorded on the traditional fishing grounds in the western North Sea. The northern distribution of the herring was in fact a serious problem during these surveys because the herring distribution apparently extended north of the survey area (the 200 m isobath). Over the deeper parts of the North Sea (more than 150 m), herring were found in the upper 50 m layer, whereas in shallower waters the herring occurred in bottom schools (Anon., 1991b).

In the last years of the decade, North Sea herring may have migrated even further out over deep water, and reached the Faroe plateau. Jacobsen (1990) reported substantial herring catches taken southeast of the Far Øer in 1990. He showed that these herring were identical in biological characteristics to North Sea herring (or the adjacent stock to the west of Scotland). This type of herring is rarely found in Faroese waters.

The question is whether the northern distribution of the herring in the late 1980s was a function of stock size and age composition, or whether is was an effect of environmental change. During the 1980s the size of the North Sea herring stock increased greatly, and also the number of older fish in the population. As older fish have a preference for deeper water, the changing age composition could have had an effect on the geographical distribution of the stock.

In order to see whether there is a relationship between stock size and distribution of the herring, the data presented in Figure 3 can be compared with estimated stock size and stock composition for the corresponding periods. The figures below have been calculated on the basis of VPA tables presented in reports of the ICES Herring Assessment Working Group (Anon., 1986, 1991a).

Period	Average stock size (2+ ringers) in '000 t	Average percentage of 4+ ringers in the adult population
1955–1960	1 403	52%
1970–1977	194	31%
1984–1987	840	31%
1988–1990	1 401	44%

It is seen that the stock in the late 1980s was neither larger nor older than the stock in the 1950s. Therefore, it is unlikely that the more northerly distribution in the late 1980s was an effect of the size or age composition of the stock.

The conclusion must be that the feeding grounds of herring in the 1980s had shifted to very unusual northeasterly positions, apparently as a result of changes in the environment. Whereas the feeding concentrations of herring were formally found in the western part of the North Sea, these concentrations in the 1980s occurred mainly in the region of the shelf edge, in the northern and northeastern parts of the North Sea.

The shelf edge along the Norwegian Trench in the 1980s not only attracted indigenous North Sea herring but also immigrant herring from the Baltic. These herring, which spawn in the southwestern Baltic in March/April, came to feed in the eastern North Sea in large numbers during the summers of 1986–1990 (Anon., 1991a). The ICES Herring Assessment Working Group estimated that most of these Baltic immigrants were distributed in summer in the area shown in Figure 4. However, the migrations of these fish may have extended even north of the area indicated. Norwegian research vessel samples in July 1990 showed spring-spawning herring to be present all the way along the Norwegian Trench up to 62°N (Anon., 1991a).

The immigration of spring-spawning herring from the Baltic into the North Sea has not been reported before 1986. It is hard to say whether it did not occur before. If it did, it must have been on a limited scale. Baltic spring spawners have a mean vertebral count that is very different from North Sea autumn spawners (55.80 vs. 56.60). Considering the traditional interest in vertebral

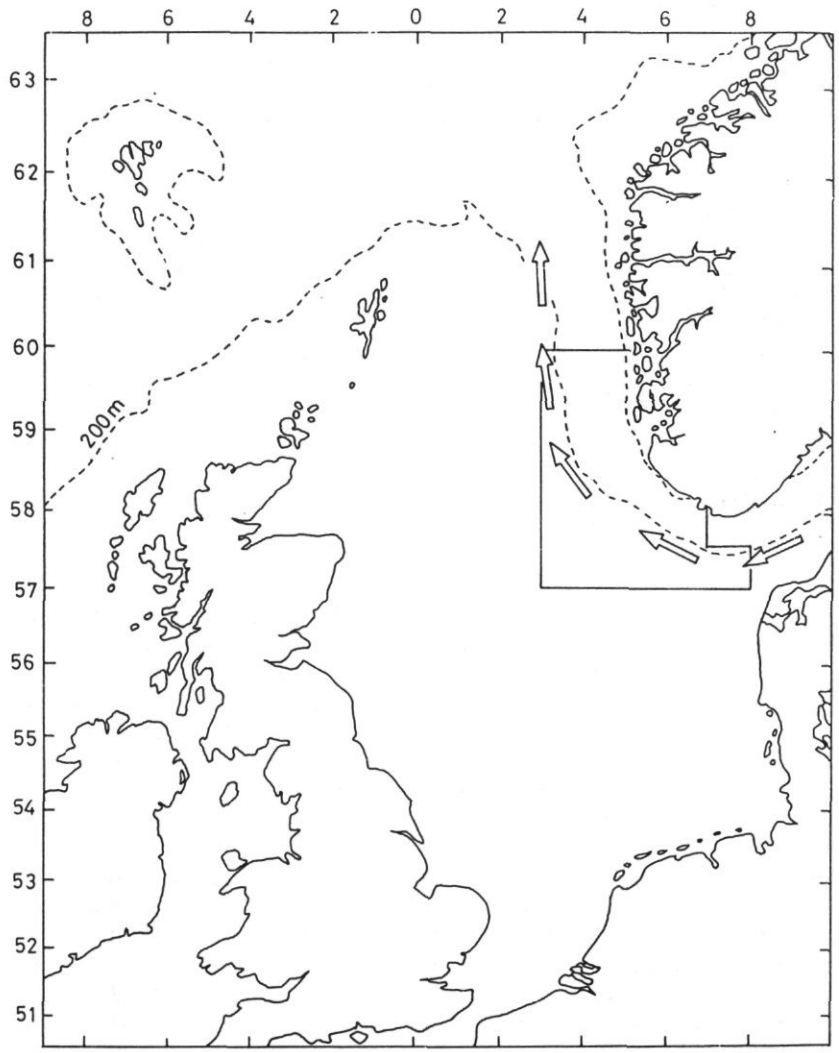

Figure 4. Immigration of Baltic herring into the North Sea during summer. The box indicates the area where Baltic herring made up a significant proportion of the catches during May–September in the years 1986–1990. Data from Anon. (1991a).

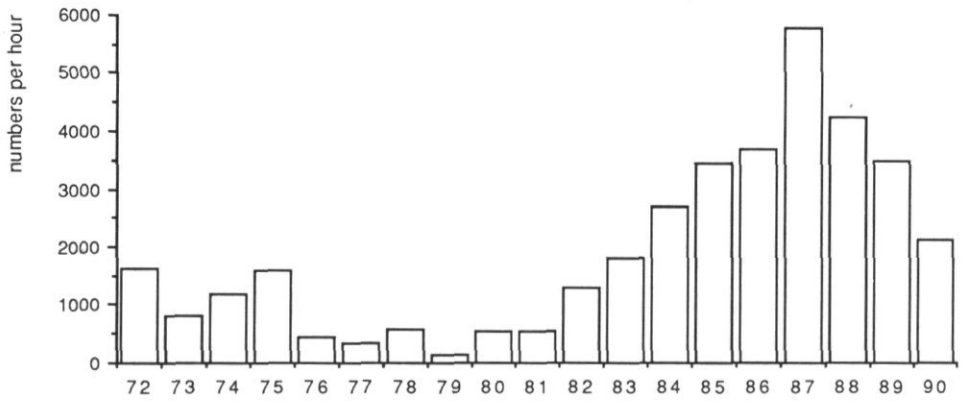

Figure 5. Abundance of 1-ringed herring in the North Sea during the International Young Fish Survey. Data from Anon. (1991a).

studies among North Sea scientists, a substantial immigration of Baltic herring into the North Sea would have been quickly detected.

Herring recruitment

After a period of extremely low recruitment in the 1970s, the strength of herring year classes gradually increased in the early 1980s, and from 1984 to 1986 the year-class strength was above the long-term average. Indices of year-class strength derived from the International Young Fish Survey are shown in Figure 5. The most striking aspect of herring recruitment in the last two decades is the apparent trend in the data. Herring recruitment did not fluctuate randomly from one year to the next, but first decreased to a very low level in the mid-1970s before very consistently increasing to reach a peak in the mid-1980s. As shown by Corten (1986), these trends in recruitment cannot be attributed exclusively to changes in parent stock. The low recruitment during the 1970s was apparently also due to an unfavourable environmental factor; a factor that gradually disappeared around 1980, and in the following years even reversed to become a favourable condition.

During the years of low recruitment, herring larvae originating from spawning grounds in the western North Sea did not reach the nursery areas in the eastern North Sea and Kattegat. For their five-month journey across the North Sea, herring larvae very much depend on the residual circulation (Fig. 7). Considering the near absence of older herring larvae from the eastern North Sea in the mid-1970s, Corten (1986) speculated that the residual currents in the North Sea had changed during this period. From 1980 onwards the abundance of older herring larvae in the eastern North Sea and Skagerrak increased, and at the same time recruitment improved.

After the herring larvae have drifted across the North Sea, they grow up either in the shallow waters of the eastern North Sea or in the Skagerrak/Kattegat area. The latter area is normally of minor importance compared to the main North Sea nurseries. However, during the period of improved recruitment, the abundance of

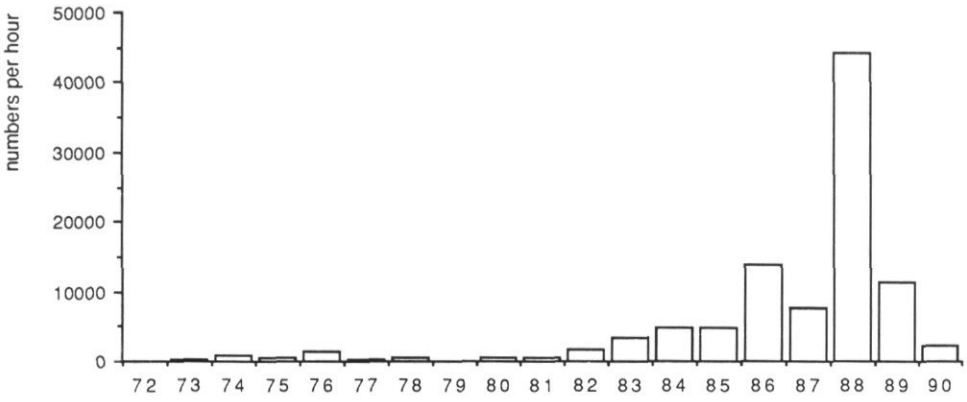

Figure 6. Abundance of 1-ringed herring of North Sea origin in the Skagerrak/Kattegat during the International Young Fish Survey. Data from Anon. (1991a).

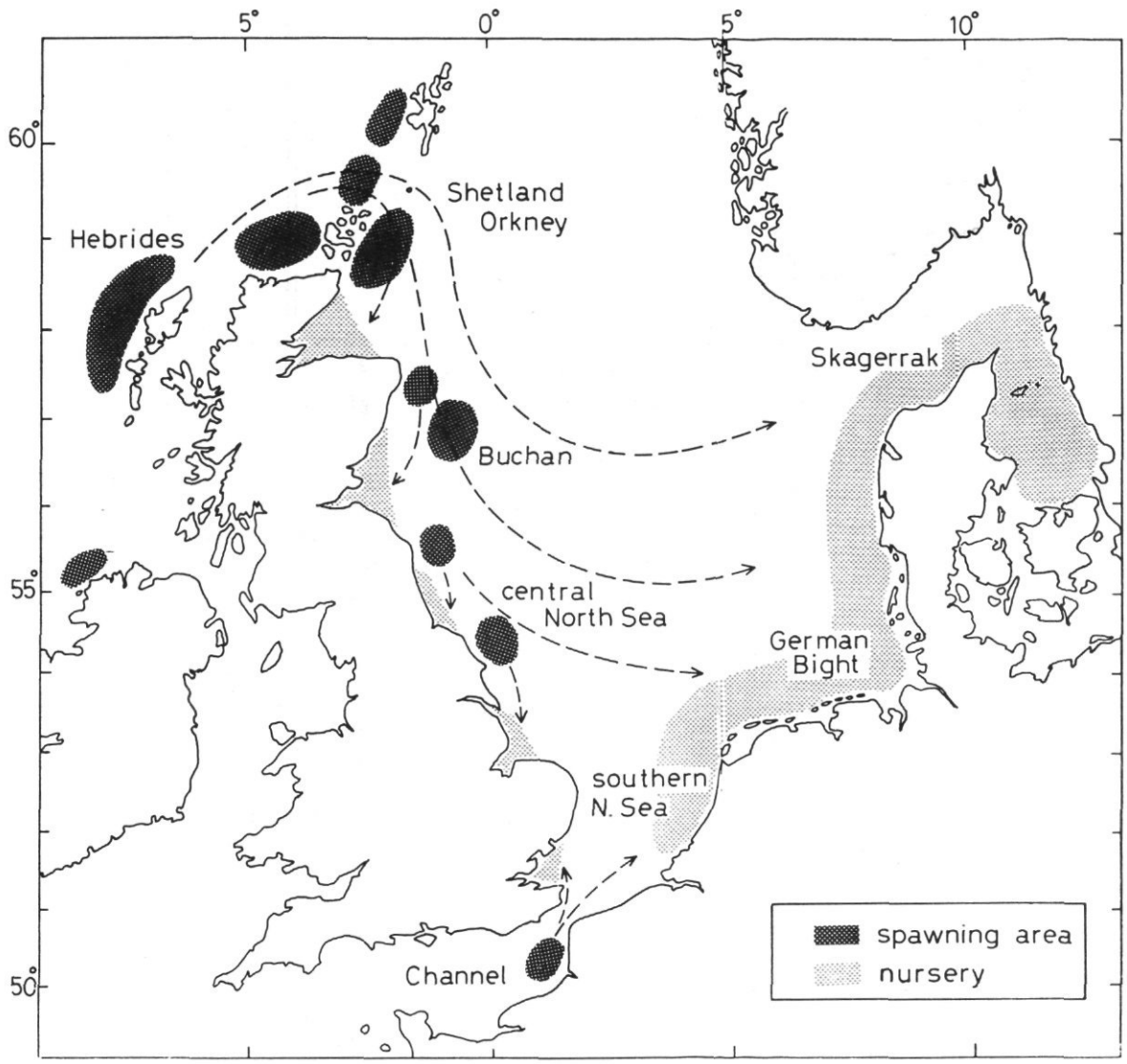

Figure 7. Drift of herring larvae across the North Sea.

juvenile North Sea herring in the Skagerrak/Kattegat increased more sharply than in the North Sea, and the relative importance of the Skagerrak/Kattegat as a nursery area for North Sea herring increased. Figure 6 shows the abundance indices of 1-ringed North Sea herring in the Skagerrak/Kattegat, and from a comparison with Figure 5 it is seen that the peak in abundance in the mid-1980s was greater in the Skagerrak/Kattegat than in the North Sea. The same fact is illustrated if for each North Sea year class the percentage of juveniles that grew up in the Skagerrak/Kattegat is calculated (Table 2). In general, the proportion of juveniles in the Skagerrak/Kattegat goes up with the absolute size of the year class.

The series of strong year classes in the mid-1980s thus coincided with an increased importance of the easternmost nurseries. One could argue that the easterly distribution of the juvenile herring was an effect of the increased size of the year classes. However, the opposite explanation might be just as plausible, namely that the absolute size of the year class was a function of the easterly distribution of the juveniles, or of the hydrographic conditions that resulted in this easterly distribution.

Changes in the distribution of mackerel

Although no new data on mackerel distribution are presented in this paper, it is interesting to review the recent literature on the subject in order to see whether

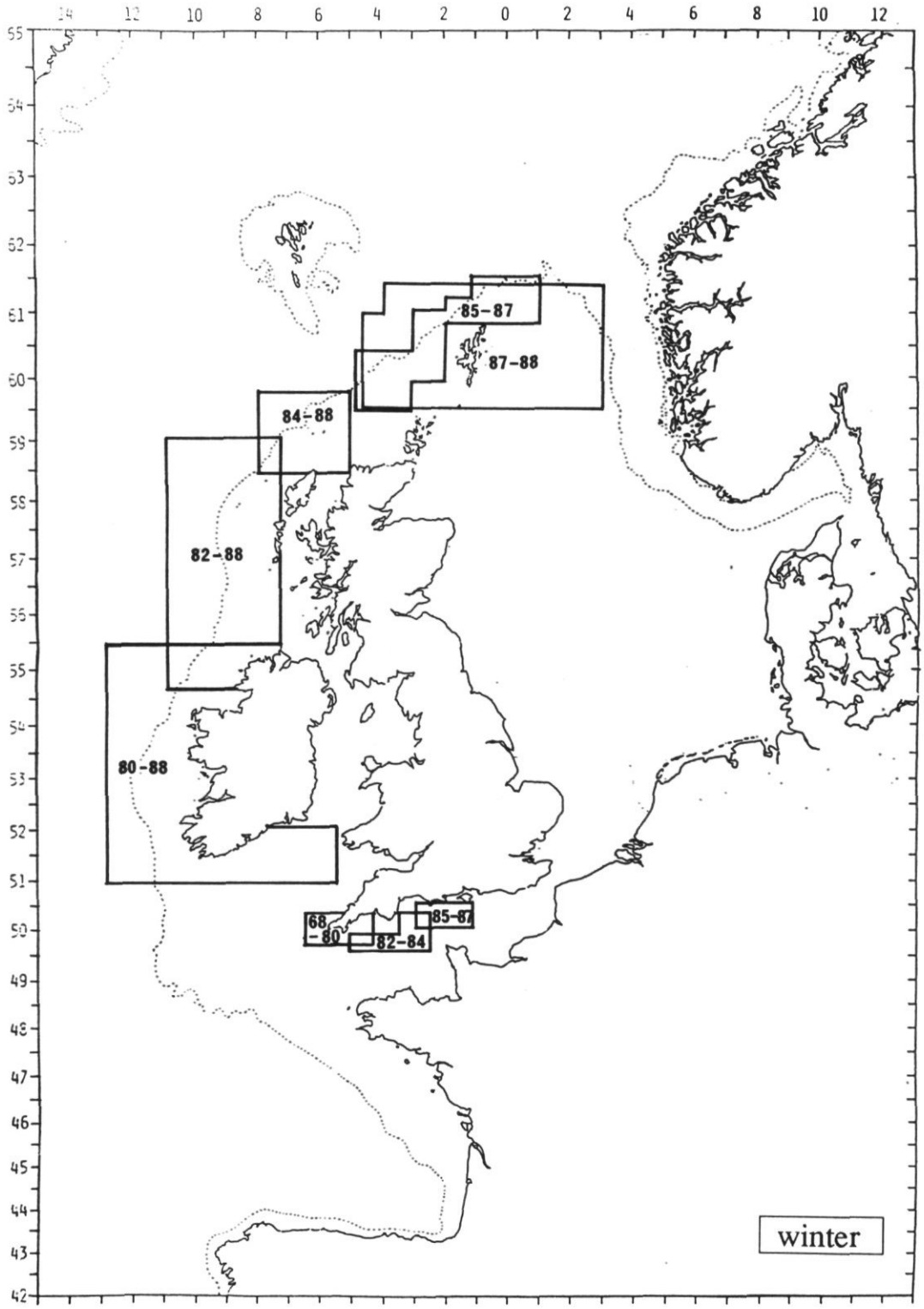

Figure 8. Shifts in the position of the main fishery on western mackerel during winter and summer. From Anon. (1990c).

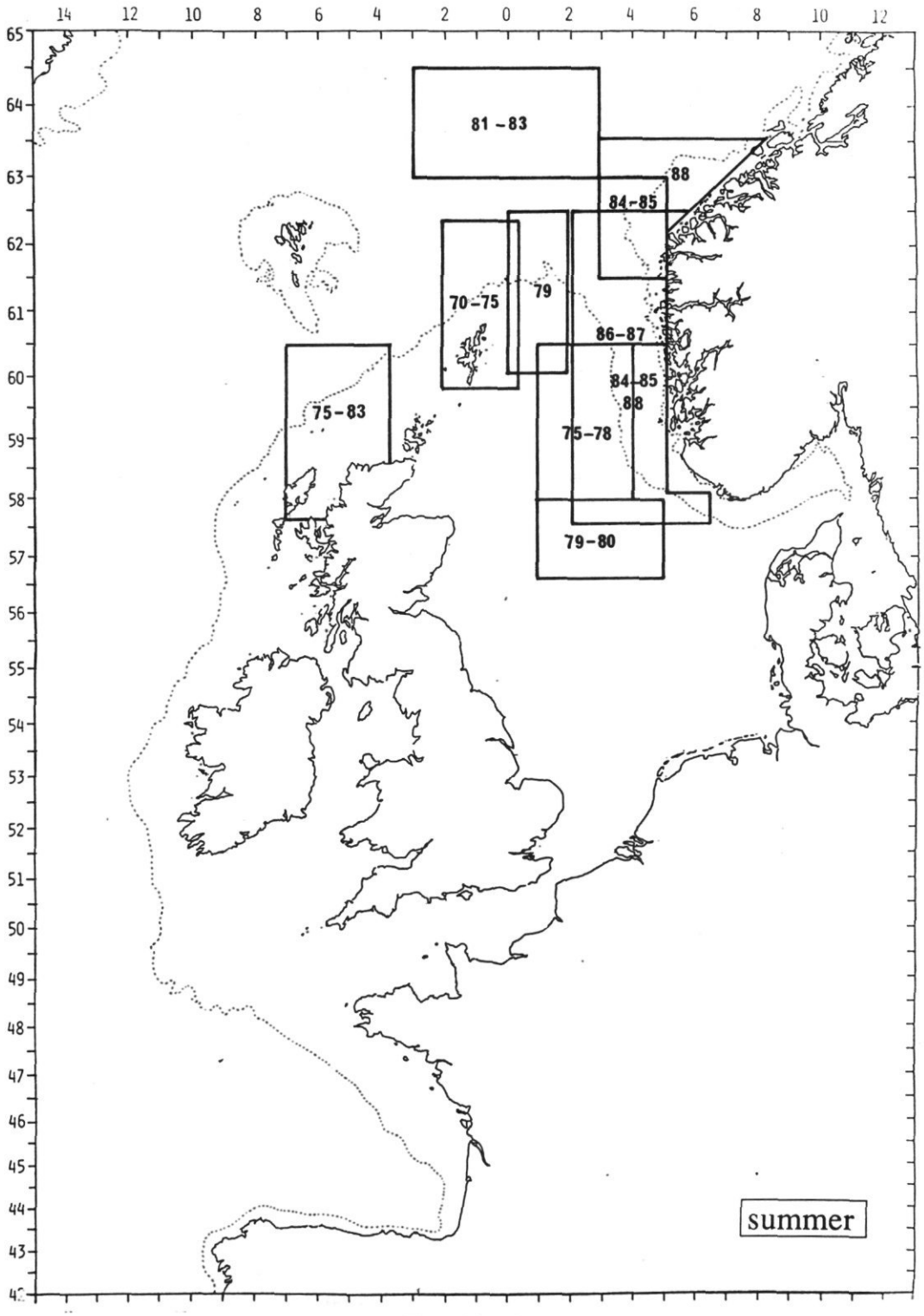

Figure 8 (*cont.*)

Table 2. Abundance indices of 1-ringed herring in North Sea and Division IIIa (Skagerrak/Kattegat). Number in each area calculated as (mean number per square) × (number of squares sampled). Data from Anon. (1990a).

Survey year	North Sea	Division IIIa	% in Division IIIa
1983	153 439	38 024	19.9
1984	163 482	31 633	16.2
1985	250 805	78 113	23.7
1986	229 255	224 370	49.5
1987	446 615	114 483	20.4
1988	259 193	661 089	71.8
1989	264 714	170 275	39.1
1990	129 486	34 583	21.1

common patterns exist in the observed changes in herring and mackerel.

The mackerel found in the North Sea in recent years predominantly belongs to the western mackerel stock (Anon., 1978). These fish have their spawning grounds along the edge of the continental shelf in the Celtic Sea and Bay of Biscay. After spawning they migrate north along the shelf edge to the waters around Shetland, and then further east into the North Sea or the Norwegian Sea. There is also a small indigenous spawning stock in the North Sea, but its size is insignificant compared to the western stock (50 000 t vs. 2 million t). The ICES Mackerel Working Group assumed that all mackerel catches taken in the North Sea in recent years (1988–1989) were derived from the western stock (Anon., 1990b).

During the last 20 years, the western mackerel has shown dramatic changes in distribution. These changes were first described in detail by the ICES Mackerel Working Group (Anon., 1985). The most conspicuous change was a shift of overwintering area from the southwest of England to the west coast of Ireland and then northward to the northwest and north of Scotland. At the same time there was also a shift in feeding area from the waters around Shetland to more eastern parts of the North Sea. In some years, particularly in the 1980s, mackerel migrated further northeast along the Norwegian coast, and crossed the border between the North Sea and the Norwegian Sea at 62°N. The changes in summer and winter distribution of western mackerel are shown in Figure 8 (from Anon., 1990c).

Walsh and Martin (1986), in providing further details on the changes in mackerel distribution, showed that the autumn fishery for mackerel west of Scotland had progressively moved further north and started later in the season. These authors were the first to speculate that the changes in mackerel distribution were a response to changes in the environment. They suggested that the northward shift in distribution since 1978 was related to an increase in North Atlantic drift along the edge of the continental shelf. This increase followed a period of apparently reduced flow in the 1970s.

Because of the political implications of the shifting distribution (the countries involved in the fishery holding different fishing rights in the various fishing zones), a special EC/Norwegian study group was set up in 1988 to describe the changes and report the conclusions to the management bodies (Anon., 1988c, 1990c).

Recent working group reports (Anon., 1990b, c) show that the trend towards a more northern and northeastern distribution of western mackerel continued in the second half of the 1980s. Since 1985 there was also an increased migration of juvenile western mackerel into the eastern North Sea and Skagerrak. The juveniles followed the same migration route as the adults; north along the 200 m isobath west of the British Isles, and then following this bottom contour into the eastern North Sea (Anon., 1990b: Figure 3.14).

In the context of the present paper, it is interesting to note in particular the changes in mackerel distribution during the feeding season in summer. Over the period 1970–1990, the summer fishery in the North Sea shifted in an easterly direction. During 1975–1980, the shift was mainly from around Shetland towards the western slope of the Norwegian Trench. In the years after 1980, the summer fishery took place not only in the region of the Norwegian Trench, but also further north along the Norwegian coast and in the Norwegian Sea.

The northern displacement of the fishery after 1980 is illustrated by a plot of catches taken north of 62°N (Fig. 9). This plot shows a steady increase in catches in the northern waters after a depression in the late 1970s.

Discussion

Natural changes in western mackerel and North Sea herring during the 1970s and 1980s have been described before (Walsh and Martin, 1986; Corten, 1986). In western mackerel, the most conspicuous change was the shift of overwintering grounds from the southwest of England to the north of Scotland. In herring, attention was focused on the dramatic changes in recruitment, and in particular on the possible changes in transport of herring larvae across the North Sea.

There was not much that the changes in both stocks had in common, other than that they happened at about the same time, and that they both happened gradually over a long period. Because of the persistent nature of the changes in herring recruitment, Corten (1986) assumed that the changes had an oceanic origin. Walsh and Martin (1986) speculated that the changes in western mackerel also had an oceanic origin, and that the northward displacement of overwintering and feeding grounds might be related to an increase in North Atlantic Drift.

The new data on herring distribution presented in this paper demonstrate a much closer connection between changes in both stocks than has been assumed until now.

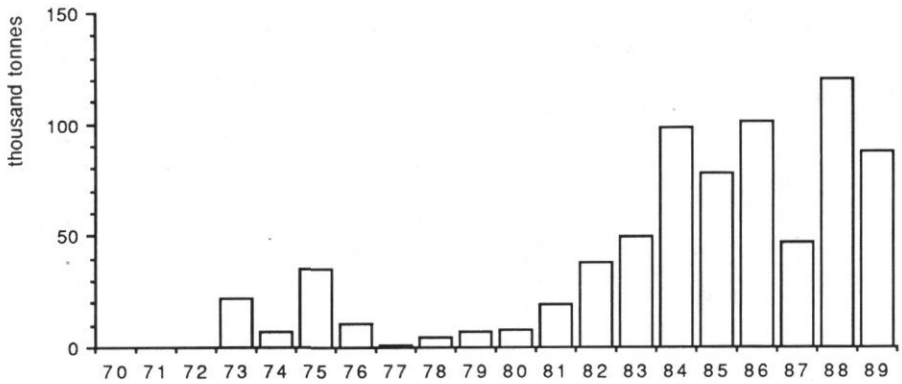

Figure 9. Catches of mackerel north of 62°N. From Anon. (1990b) and earlier reports of the Mackerel Working Group.

The shift towards feeding grounds in the northeastern North Sea, documented earlier for western mackerel, is now also seen in North Sea herring. This strongly suggests that both stocks are influenced by the same oceanic factor.

It is not yet clear what the nature of this oceanic factor has been. A change may have occurred either in the transport of oceanic water in the northern region of the North Sea, or in the chemical or physical properties of the water in this area. As already mentioned, Walsh and Martin (1986) assumed that the hydrographic factor that affected mackerel distribution was a change in water flow along the edge of the continental shelf.

The present authors believe that such a change in water transport would also provide the best explanation for the observed changes in herring distribution. There is a remarkable coincidence between the position of herring catches during the later part of the 1980s and the position of the shelf edge current (Fig. 10). This current, which consists of a relatively narrow core of saline water, flows northward along the edge of the continental shelf of western Europe (Dooley, 1969, 1974). It follows the 200 m isobath in the northern North Sea until the entrance of the Norwegian Trench. At this point it branches into two currents: one that flows northeast, following the 200 m isobath along the Norwegian coast, and another that turns south along the western edge of the Norwegian Trench. This is the Tampen Bank current, named after the position where it branches off the main shelf edge current. Because the saline water of the Tampen Bank current is heavier than the relatively fresh water from the Baltic outflow, the Tampen Bank current sinks to the bottom, but continues its course along the shelf edge in the direction of the Skagerrak.

The distribution of herring catches in June/July during the later years of the 1980s was very much concentrated in the region of the shelf edge current, and in particular along its southern branch, the Tampen Bank current. The fact that herring showed an unusual concentration in this area suggests that something has happened to the shelf edge current.

If we compare the changes in summer distribution between herring and mackerel, we see that both species shifted to the northeast during the 1980s. Herring was found mainly along the northern edge of the North Sea and along the Norwegian Trench, whereas mackerel, too, was found further north along the Norwegian coast. What these areas have in common is that they are all influenced by the shelf edge current.

To explain the northward shift in mackerel, Walsh and Martin (1986) postulated an increase in strength of the shelf edge current. The western mackerel spend most of their life inside this current or in its immediate vicinity, and it is possible that an intensification of the

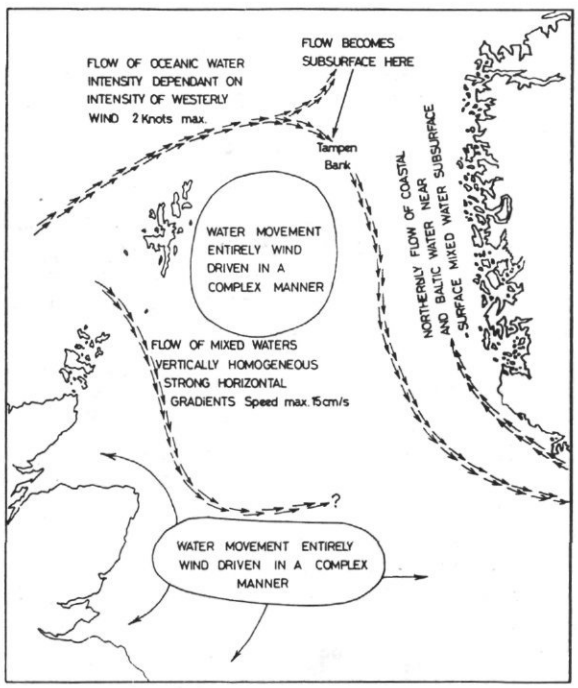

Figure 10. Residual currents in the North Sea. From Dooley (1974).

current would automatically lead to a northward displacement of the stock.

For North Sea herring, such a simple explanation is not possible. These herring normally spend most of their lives in the shallower part of the North Sea, and must have had a special reason for concentrating in the region of the shelf edge current during the 1980s. One theory explaining the sudden attractiveness of the shelf edge for North Sea herring is that an intensification of the shelf edge current during the 1980s resulted in an increased vertical mixing in this area, and a subsequent increase in biological productivity. This would explain why the herring was concentrated in this area precisely during the feeding season. This hypothesis, however, has to be tested by considering long-term series of plankton data for the shelf edge area.

At the same time when changes occurred at the shelf edge of the North Sea, other changes may have taken place in more southern parts of the North Sea. It is not clear whether herring left the former feeding grounds in the western North Sea because of the development of richer feeding areas elsewhere, or because of a decline in absolute food production in the western North Sea. In the latter case (which seems the more plausible), changes may have occurred in the quantity or quality of Atlantic inflow in the northwestern North Sea.

The changes in herring recruitment in the 1970s and 1980s also suggest that long-term changes took place in residual currents in the western and central North Sea. The poor survival of herring year classes in the period 1972–1978 was ascribed by Corten (1986) to a reduction in transport of larvae from the western North Sea to the nursery areas in the eastern North Sea and the Skagerrak/Kattegat. In the 1980s, the opposite situation may have occurred. At the time when herring recruitment increased to a level above average, the year classes at the age of 1.5 years had a relatively easterly distribution. The prolonged period of strong recruitment in herring may thus have been related to a period of consistently increased eastward transport across the central North Sea.

It is worth noting that pronounced long-term cycles have earlier been described for herring stocks in waters adjacent to the northern North Sea. Both the Atlanto-Scandian herring spawning along the Norwegian west coast and the Bohuslan herring along the Swedish Skagerrak coast have shown a long-term cycle of rise and decline, and the cycles in both stocks seem to be connected (Devold, 1963; Höglund, 1972). In both cases, the areas where long-term changes occurred are under the direct influence of the North Atlantic.

The changes in North Sea herring and Western mackerel described in this paper also occurred on the border of the North Sea and the Atlantic Ocean. It is possible, therefore, that the changes in these stocks are somehow related to the well known cycles in Atlanto-Scandian and Skagerrak herring.

Acknowledgements

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The stock of Atlanto-Iberian sardine: possible causes of variability[1]

R. Robles, C. Porteiro, and J. M. Cabanas

Robles, R., Porteiro, C., and Cabanas, J. M. 1992. The stock of Atlanto-Iberian sardine: possible causes of variability. ICES mar. Sci. Symp., 195: 418–423.

Fluctuations in the Iberian stock in the period 1980–1989 have been examined, taking into account what is known of the biology of the species, oceanic and atmospheric parameters, and the fishing effort directed at this stock. During this period, there have been significant changes in recruitment, which was good in 1983 and 1987 and poor in 1982, 1985, 1986, and 1988. Some links between recruitment anomalies and environmental parameters (surface temperatures, prevailing winds, sea level) have been identified. Variations in recruitment can be explained on the basis of fluctuations in "environmental windows" during the periods of spawning and larval drift. The success of spawning is controlled by the existence of a favourable environment during larval drift into areas with weak upwelling and abundant food.

R. Robles, C. Porteiro, and J. M. Cabanas: Instituto Español de Oceanografía. PO Box 1552, 36280 Vigo, Spain.

Introduction

The variability of recruitment in pelagic fish populations is an important problem in fisheries science. Traditional stock-recruitment models are based exclusively on the integration of biological interactions during the pre-recruit phase, but they ignore abiotic processes during recruitment, interspecific interactions, and the effects of climatic change on the habitat.

The mechanisms which govern large-scale spatial changes are not well understood, but in coastal systems, short and medium-term fluctuations in the abundance, distribution, and production of the ecosystem have been related to large and mesoscale variations in atmospheric processes which have an enormous influence on the upper ocean (Lasker, 1978).

On decadal time scales, environmental fluctuations cause variability which must be amplified by exploitation. Rational exploitation could smooth the natural fluctuations of the stock and attain an equilibrium between fishing and production; nevertheless, environmental fluctuations can cause large and prolonged changes in the mean size of the stock, i.e. prolonged periods of high or low biomass.

The sardine (*Sardina pilchardus*, Walb.) is the target of an important fishery on the Atlantic coast of the Iberian peninsula. Its capture and associated socio-economic infrastructure indicate that it may be one of the most ancient fisheries in this region.

The sardine belongs to a low trophic level. The biological characteristics show that its populations are well adapted to environmental fluctuations (Margalef, 1960). The northern and southern limits of the global distribution of the European sardine are well defined, in the north by the 10°C surface isotherm, and in the south by the 20°C isotherm. In the northern part (France, British Isles) its distribution overlaps that of the herring (*Clupea harengus*), and in the south (West Africa, eastern Mediterranean) with the alache (*Sardinella aurita*) (Wyatt, 1985).

On account of the broad distribution of this species, there are considerable variations in the life cycle characteristics, but here we refer only to studies of the sardine stock of the Atlantic waters of the Iberian peninsula. We assume for present purposes that there is a single stock distributed between France and Gibraltar.

Geographically this area lies at the northern limit of the east central Atlantic coastal upwelling system (Wooster *et al.*, 1976), which in these latitudes is seasonal (April to September) and relatively weak (Fraga, 1981; Blanton *et al.*, 1984; Fiuza, 1984).

[1]To the late Reuben Lasker and to Warren Wooster, wholehearted supporters since 1985 of the work by the Instituto Español de Oceanografía in recent years to further our knowledge of factors which affect the variability of recruitment in coastal pelagic fish.

Two regions can be distinguished in the Iberian peninsula from an oceanographic viewpoint, the Cantabrian Sea and the Atlantic, clearly differentiated and separated by a convergence which lies between Cape Ortegal and Cape Finisterre at the corner of the peninsula. Within each region it is possible to identify two slightly different sub-regions which may influence the behaviour of pelagic species (Fig. 1).

Two areas, eastern and western, can be distinguished on the Cantabrian shelf in the period March to October, a period important for the survival of larvae and pre-recruits (areas 1 and 2 in Fig. 1), with different dynamic and thermal features. Surface flow is to the east in the eastern area and can form gyres; the water here is warmer (mean temperature 2°C higher than elsewhere on the platform). Surface flow is to the west in the western area due to the predominance of northerly winds at this time, with zones of weaker circulation, and again a tendency to form gyres (Servain, 1976; Ibañez, 1984).

The west Atlantic coast (areas 3 and 4 in Fig. 1) is subject to seasonal upwelling and surface currents parallel to the coast in the direction of the dominant winds, with a marked northerly component (Cabanas *et al.*, 1988; Coste *et al.*, 1986; Fiuza, 1984). It should be noted that in the northern part (area 3, very sinuous, indented by rias) there is a convergent zone close to Cape Finisterre which affects the exchange of water between the Cantabrian platform and the Atlantic, and deflects some offshore (McClain *et al.*, 1986).

The reproduction and feeding periods of the sardine seem to be adapted to the conditions of each area, and although sardines are found in all areas, they tend to be distributed by age classes (Porteiro *et al.*, 1986). The 0-group fish (recruits) are found mainly on the west coast of the Iberian peninsula between Cape Ortegal and Cape Roca, the zone of maximum upwelling at the time when the recruits reach this area (Dias *et al.*, 1983; Pastor *et al.*, 1986; Dias *et al.*, 1989). The 1-group fish occur mainly in the southern part of the area occupied by the stock, in Portuguese waters. Fish of five years and older are found in the Cantabrian Sea (Pastor *et al.*, 1986; Porteiro *et al.*, 1990).

The Iberian sardine has two spawning areas, off the Portuguese coast in winter (Cunha and Figueiredo, 1988) and in the Cantabrian Sea in spring (García *et al.*, 1988). On the north Portuguese coast, maximum spawning takes place in December at temperatures between 14°C and 16°C (Ré *et al.*, 1990). In the Cantabrian Sea, spawning takes place between October and July, with maximum intensity in April and May, and a smaller peak in November and December, the maxima occur at temperatures between 12.5°C and 15°C (Chesney and Alonso, 1988; Solá *et al.*, 1990).

Analysis of sardine stock trends during the period 1976 to 1989 (Pestana, 1989) shows that in the short term the yield depends on the recruitment indices, as is characteristic of short-lived pelagic species (Ulltang, 1980).

In this work, an attempt is made to qualitatively relate the variability of certain environmental factors to the variations observed in recruitment, and thus to changes in sardine stock abundance.

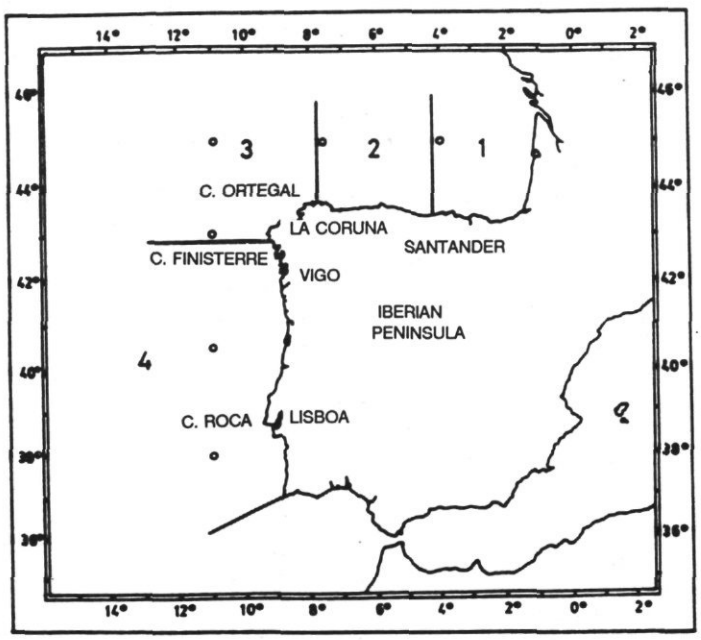

Figure 1. Situation map, with indications of zones and points where upwelling indices are calculated.

Table 1. Recruitment, landing of juveniles, upwelling index.

Year	Recruitment (*)	Upwelling index (**)	Landing juveniles (*)
1980	26 784	188	388
1981	17 709	337	329
1982	12 466	193	2
1983	35 468	13	1 179
1984	14 990	456	157
1985	11 530	118	21
1986	9 506	470	11
1987	18 547	241	788
1988	9 943	245	9
1989	9 272	379	95

* millions of individuals.
** m^3/s km.

Methods

Sardine stocks are assessed by the ICES working group using virtual population analysis, calibrated by acoustic surveys and the catch per unit effort of the fishery. From the age-class distribution the abundance of 0-group fish is obtained, which corresponds to recruitment (Table 1) (Anon., 1990).

The monthly captures of juveniles (length range 8–12 cm) landed in southern Galician ports (Table 2 and Fig. 2) are correlated ($r = 0.8$) with the VPA recruitment estimates. By assuming that juveniles are caught six months after hatching (based on studies of the daily growth rings of the otoliths, (Alvarez and Butler, 1989)), it is possible to determine the most successful spawning period each year (Table 3).

Two series used to examine the environmental variability during this period are the upwelling indices calculated from the wind stress on the sea surface (Bakun, 1973), and mean sea level in the port of Vigo (Lavín and García, 1991). The upwelling indices ($-Mx$, Ekman mass transport in the x-axis) gives us an idea of surface transport due to the wind. Upwelling provides nutrient-rich water to the euphotic zone. Sea level is also related to the wind and other coastal oceanographic phenomena.

The daily upwelling indices along the coast are estimates using the geostrophic winds (Lavin *et al.*, 1991). It can be seen from the 1988 series that the station at 43°N, 11°W was representative of the whole area. Correlation analyses give $r>0.8$ with each of the other stations (Fig. 1). Table 1 gives the mean upwelling indices during the active period of upwelling (April to September) for the representative station.

From the wind data, we can extract the periods of five consecutive days when wind speeds were less than 6 m/s; these periods, called "Environmental Windows", are periods when turbulence and food availability as larval growth limiting factors in upwelling areas are minimized; and may be one of the keys to recruitment success (Cury and Roy, 1989). Table 4 lists the monthly presence (+) or absence (0) of such periods.

Results and discussion

The correlations between recruitment and the upwelling index for April to September (active upwelling season) and March to May (spawning season) are similar ($-0.5 < r < -0.6$) for the period 1976 to 1989 (Cabanas *et al.*, 1989) and for the period 1980 to 1989. Although they are not significant, upwelling appears to influence annual recruitment. But this index cannot on its own be considered a good forecast of recruitment. In part forecasting may not be possible without separating the pre-recruit phase into at least two stages, the larval drift stage when transport plays an important role, and the post-larval stage when survival depends on the abundance and availability of food.

Previous studies have tried to find the causes of sardine fluctuations in this upwelling area by comparing them with the position and strength of the Azores anticyclone. Dickson *et al.* (1985) found a negative correlation between the upwelling indices and sardine landings in area IXa for the period 1950 to 1984, and concluded that an increase in the intensity of northerly winds increased turbulence and Ekman transport, and reduced production at all levels in the food chain. Fiuza *et al.* (1982) found a positive correlation between the

Table 2. Landing of juveniles in the Galician zone (t).

	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989
Jan	559	761	223	0	170	0	0	0	17	23
Feb	17	145	112	0	1 802	0	0	0	136	1
Mar	119	33	2	0	454	0	0	0	35	0
Apr	33	61	0	0	1 788	0	0	0	23	0
May	107	1	0	0	1 488	0	0	0	24	0
Jun	136	0	0	0	861	0	0	117	0	0
Jul	40	0	0	2	899	0	0	1 258	0	46
Aug	259	4	0	87	1 138	0	16	1 689	0	252
Sep	1 306	238	17	1 335	0	19	3	3 115	0	607
Oct	1 067	980	0	3 279	0	118	10	1 456	0	196
Nov	968	2 660	5	2 768	0	121	0	1 970	117	69
Dec	958	252	0	1 290	0	18	0	527	1	1

Table 3. Percentage of stock contribution according to birthdate.

	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989
Jan	–	–	–	–	2	–	–	4	–	–
Feb	–	–	–	–	3	–	–	5	–	1
Mar	1	1	–	4	–	–	–	8	–	1
Apr	6	3	–	8	–	–	–	4	–	1
May	3	6	–	7	–	1	–	5	–	–
Jun	1	1	–	3	–	–	–	1	–	–
Jul	2	–	–	1	–	–	–	–	–	–
Aug	–	–	–	5	–	–	–	–	–	–
Sep	–	–	–	1	–	–	–	–	–	–
Oct	–	–	–	5	–	–	–	–	–	–
Nov	–	–	–	4	–	–	–	–	–	–
Dec	–	–	–	2	–	–	–	–	–	–
	13	11	0	40	5	1	0	27	0	3

monthly percentage of the total Portuguese landings (1939 to 1974) and upwelling indices three months earlier. Wyatt and Pérez-Gándaras (1986) also found a positive correlation between annual captures and dominant winds on the western shelf of the Iberian peninsula.

Table 3 shows that the period March to May is the most important part of the spawning season in terms of production of recruits. It can also be seen that there are periods when spawning produces few recruits, perhaps because the environmental conditions are unfavourable for larval survival. So any attempt to explain relations between environmental conditions and recruitment must take account of both upwelling indices and the environmental window concept.

From observations summarized in Tables 3 and 4 on successful spawnings and environmental windows, it can be deduced that in the months with a high survival of brood there was always an environmental window. Tables 3 and 4 show that in the period April to July, when there are no windows (0), the production of recruits is very small or fails completely. On the other hand, the more successful spawnings correspond to years with "Environmental Windows" in the months following the spawning, indicating periods of low turbulence and weak upwelling on the west coast.

In the first stage the coastal current must aid the transport of larvae from the spawning zone (Cantabrian Sea) to the pre-recruit zone in the West (Galicia). In this period (April to June) the circulation changes from the winter to the spring–summer pattern, with a change in the direction of the surface current. Depending on the timing of the change and on the velocity of the current, the brood from the Cantabrian Sea will either be transported to the West Galician coast where there is abundant food associated with coastal upwelling, or it will be dispersed off Cape Finisterre towards the open sea where many larvae die of starvation. But some larvae survive this obstacle, and reach the nursery area on the Atlantic coast, an upwelling zone, where food is abundant.

In 1983 and 1987, sea level on the west coast (Vigo) was significantly higher than during the rest of the decade, while in 1981, 1982, and 1986 sea level was lower than normal. The 1983 year class was the strongest on record, and the 1987 year class was also relatively strong. The 1988, 1989 year classes were the poorest in the decade (Fig. 3a).

Recruitment was poor in 1982, 1985, 1986, and 1988, perhaps due to dispersal of the eggs or transport to the open sea during periods of larval abundance, coinciding with the absence of environmental windows. Recruitment was very good in 1983 and 1987, years with en-

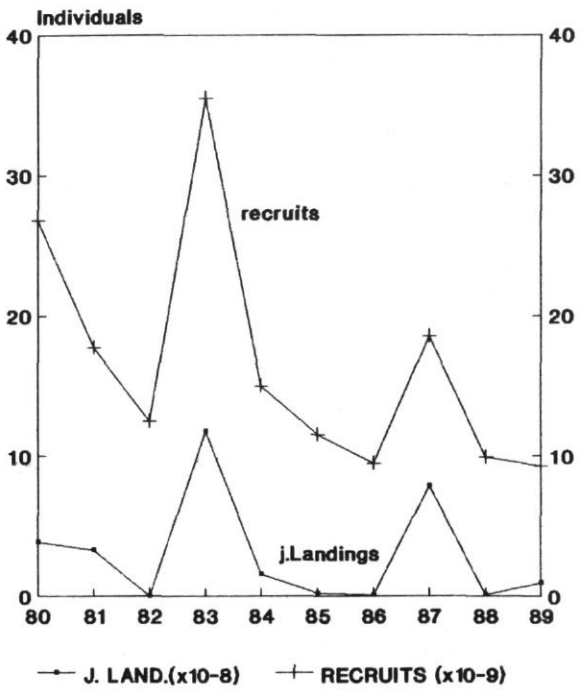

Figure 2. Landings of Juveniles–Recruits.

Table 4. Environmental windows (5 days consecutively of wind speed less than 6 m/s).

	1980	1981	1982	1983	1984	1985	1986	1987	1988	1989
Jan	0	+	+	+	0	0	+	+	0	+
Feb	0	+	0	+	+	+	0	+	+	+
Mar	+	0	+	+	+	+	0	+	0	+
Apr	+	+	0	+	+	0	0	+	+	0
May	+	+	+	+	0	+	+	+	+	+
Jun	+	+	+	+	+	0	0	+	0	0
Jul	+	+	+	+	+	+	0	+	+	+
Aug	+	+	+	+	+	+	+	+	+	+
Sep	+	+	+	+	+	+	+	+	+	0
Oct	0	+	0	+	+	+	+	+	+	0
Nov	0	+	+	+	0	0	0	0	0	0
Dec	0	+	+	+	+	0	+	0	+	0

+: presence of Environmental Windows.
0: absence of Environmental Windows.

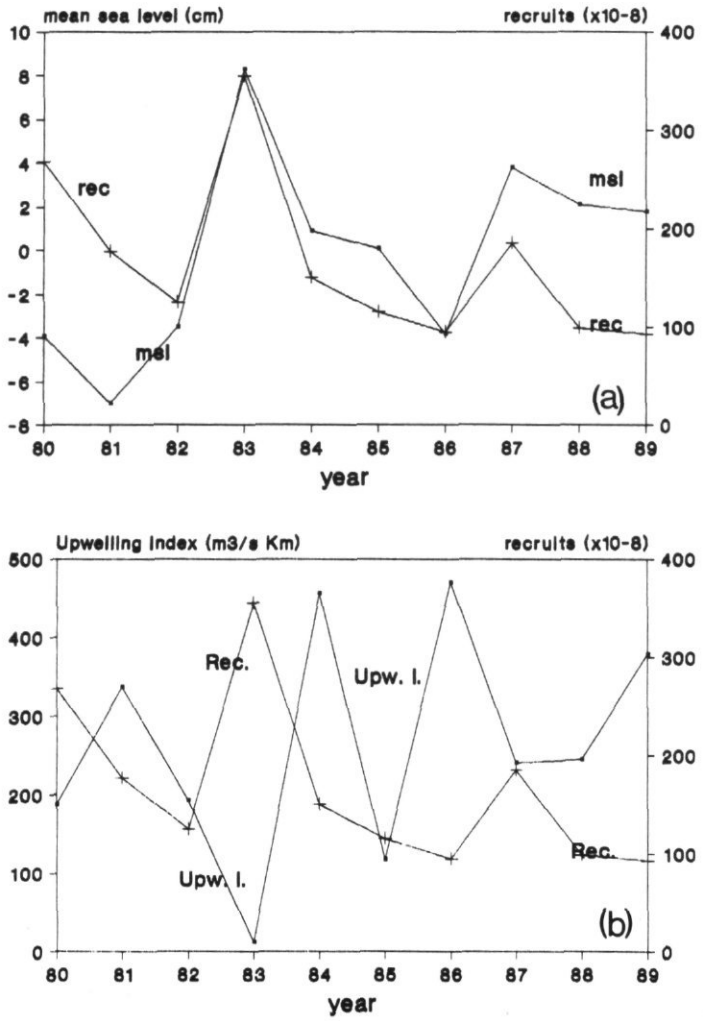

Figure 3. a. Mean sea level anomaly (Vigo) and recruitment. b. Upwelling index and recruitment.

vironmental windows throughout the larval phase. During this decade, the stock has varied within its normal limits, considering the cyclical oscillations of pelagic stocks, and is now in a period of low recruitment.

No clear trends have been identified during this decade in environmental parameters (upwelling index, temperature, sea level) which have remained within normal limits (Fig. 3). Only during 1983 were significant climatic anomalies noted; the upwelling season was reduced to half (to three months), and sea level was higher throughout the year than during the rest of the series.

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Variation in the distribution of fish stocks on the northeast continental shelf in relation to their environment, 1980–1989

D. G. Mountain and S. A. Murawski

Mountain, D. G., and Murawski, S. A. 1992. Variation in the distribution of fish stocks on the northeast continental shelf in relation to their environment, 1980–1989. – ICES mar. Sci. Symp., 195: 424–432.

Time-series observations of fish distributions and environmental conditions are used to investigate environmentally induced variation in the fish stocks on the northeast continental shelf of the USA. Standardized bottom-trawl surveys, which have been conducted each autumn since 1963 and each spring since 1968, allow the conditions within the decade of the 1980s to be compared with those of the two previous decades.

Significant temperature changes occurred in the spring during the period of these observations, with the 1980s being intermediate between the colder 1960s and warmer 1970s. Interannual variations of ± 2–4°C were observed in all shelf regions. Large decreases in the abundance of many stocks caused by heavy fishing pressure confound analyses of the effects of environmental influence on stock abundance. Changes in the distribution of selected stocks, however, can be identified and attributed to changes in temperature. The distributional change may appear either as a north–south shift or as a change in water depth. The range of observed temperature change is comparable to that anticipated in the Northwest Atlantic under a CO_2 scenario. These results may be used as an indicator of potential response of the fish stocks to climate change.

D. G. Mountain and S. A. Murawski: National Marine Fisheries Service, Northeast Fisheries Center, Woods Hole, Massachusetts 02543, USA.

Introduction

Identifying the influence of environmental variability on fish stocks is both a desirable and difficult task (e.g. Shepherd *et al.*, 1984). A changing environment may affect stock abundance (through variable recruitment) and/or stock distribution. Both of these stock characteristics also may be confounded by the influence of fishing on stock abundance, making it difficult to confidently identify those changes due to environmental variability.

In the Northwest Atlantic, analyses have indicated large-scale coherence in recruitment patterns of different fish stocks and suggested significant relationships to selected climatic variables (Koslow, 1984; Koslow *et al.*, 1987). Owing to the non-stationary nature of the recruitment data, however, longer time series will be required to confirm these relationships (Thompson and Paige, 1989; Cohen *et al.*, 1991). Other analyses have documented changes in the distribution of a number of fish stocks in response to interannual changes in water temperature (Anderson and Almeida, 1977: Murawski and Mountain, 1990).

In this report, changes in the abundance and distribution of finfish stocks on the northeast continental shelf of the USA are compared with changes in the temperature of the shelf waters. The time period considered is 1968 to 1989, with particular reference to changes during the 1980s. The response of the fish species in the shelf region to these temperature changes could be an important indication of the possible response of the fisheries to climate change predicted for the Northwest Atlantic (Murawski and Mountain, 1990).

Data and methods

The Northeast Fisheries Center (NEFC) has conducted standardized bottom trawl surveys each fall since 1963 and each spring since 1968 (Azarovitz, 1981). A stratified random survey design is used, with approximately 300 stations being occupied on a survey (Fig. 1). For comparing the fish catch data and water temperatures (surface and bottom) from these surveys, only the spring data are used. In the spring, after winter cooling and mixing, the water column is relatively uniform in temperature. In the autumn the water column is thermally stratified, with surface to bottom temperature differences of up to 10°C, making it difficult to associate the fish catch at a station with a particular temperature.

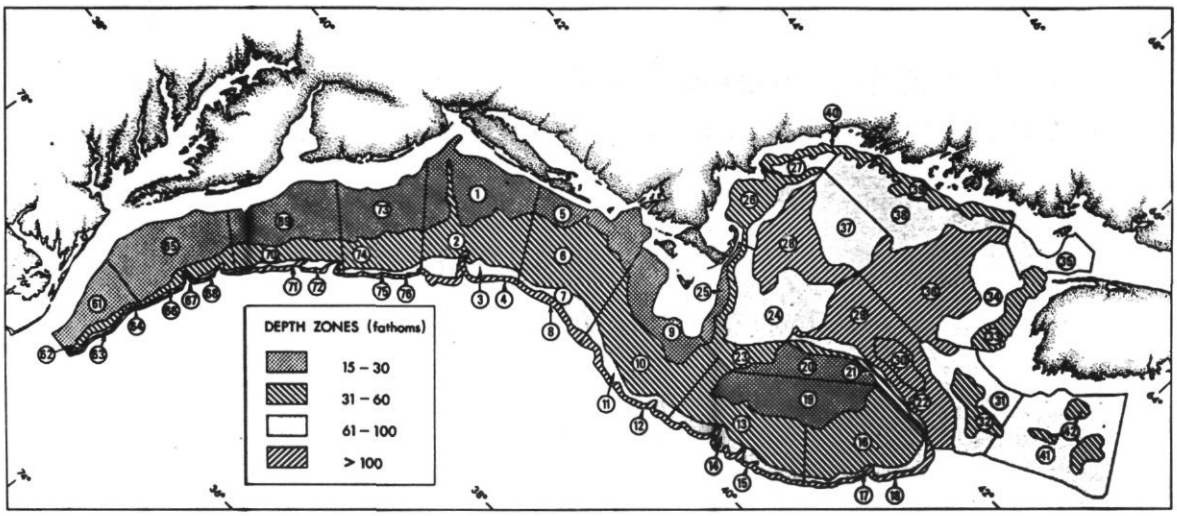

Figure 1. Offshore bottom trawl survey strata off the northeast USA, used in conducting stratified trawl surveys. Survey station allocation is generally proportional to stratum area.

Following the methods outlined by Murawski and Mountain (1990), the weighted mean bottom depth, surface temperature, bottom temperature, and latitude of capture have been calculated for each of 30 species of finfish on each spring survey (Table 1). The weighting was done by the $\log_e (n + 1)$ of the number of a species caught at each station on a survey. The mean log abundance of each species was also calculated for each survey to represent the change in relative abundance during the trawl survey time series. In doing the calculations, no separation was made between juveniles and adults – all sizes (and therefore all ages) in the trawl catch data were included.

Mountain and Holzwarth (1990) calculated the areal average surface and bottom water temperature for different regions of the northeast shelf from the temperature data collected on the trawl survey cruises. These values have been combined to provide areal average temperature values for the whole shelf and for three major subregions: the Gulf of Maine, Georges Bank, and the Middle Atlantic Bight (MAB).

The relationship between fish distribution and water temperature conditions was investigated with a general linear modeling approach. For each of the three catch variables, the mean temperature, mean bottom depth, and mean latitude, a regression model was determined using the regional average temperature and the mean abundance as independent variables. The areal average temperature for the region of the shelf where each species is characteristically located in the spring was used in the analysis. For pelagic species, surface temperature was used and for demersal species, bottom temperature was used. The mean abundance was included in the regression analysis to test if changes in stock size contributed to the changes in the three catch variables – which could occur if a species were preferentially depleted in one region of the shelf.

Results

Temperature: The temperature conditions observed on the NEFC trawl surveys have been summarized by Mountain and Holzwarth (1990), who reported surface and bottom temperature anomalies for each survey, relative to a reference temperature cycle derived from a separate hydrographic data set for the years 1977–1987. The anomaly calculation compensates for variability in the timing of the surveys in different years. The spring anomalies for the whole northeast shelf, the MAB, Georges Bank, and the Gulf of Maine are shown in Figure 2. The anomaly series are correlated significantly between surface and bottom values and between the different regions. A range of about 3°C was observed for the shelf as a whole, with larger variation in the MAB and smaller in the Gulf of Maine. The anomalies series are autocorrelated with multiple year periods of generally warmer and cooler conditions. The late 1960s was the coolest period, the mid-1970s the warmest, while the late 1970s and early 1980s were intermediate between the previous periods. The mid-1980s was again warm, although not as warm as the mid-1970s.

Abundance: Indices of abundance for different species groups on the northeast shelf are presented in Figure 3 (from Conservation and Utilization Division, 1991). During the 1980s the principal groundfish, flounders, and other finfish declined significantly in abundance from the modest increase in biomass which occurred in the late 1970s. This continued decline in abundance has been due to fishing mortality (Conservation and Utilization Division, 1991). The continued

Table 1. Results of regression analyses with areal average spring temperature (Temp) and stock abundance (Abund) as independent variables and the weighted mean catch parameters – catch temperature, catch bottom depth, and catch latitude – as dependent variables. The coefficients for Temp and Abund are listed for regressions with probability <0.05. The areal average temperature used for each regression is indicated by S or B for surface or bottom temperature and by area, the whole shelf (shelf), Middle Atlantic Bight (MAB), Georges Bank (GB), or the Gulf of Maine (GOM). For the catch temperature regressions an * indicates that the slope of the regression was not significantly different from 1.

			Catch temperature			Catch depth					Catch latitude				
Species	S/B	Area	R_2	Prob	Temp	R_2	Prob	Temp	Prob	Abund	R_2	Prob	Temp	Prob	Abund
Spiny dogfish	B	shelf	0.31	<0.01	0.50	0.23	0.02	−8.2			0.33	<0.01	0.36		
Winter skate	B	MAB	0.33	<0.01	0.41						0.29			<0.01	−0.50
Little skate	B	MAB	0.64	<0.01	0.63	0.19			0.04	−12.5	0.18			0.05	−0.41
Atlantic herring	S	MAB	0.37	<0.01	0.44						0.60	<0.01	0.49	<0.01	−2.28
Alewife	S	MAB	0.38	<0.01	0.48						0.25	0.02	0.29		
Blueback herring	S	MAB									0.26	0.02	0.41		
Silver hake	B	shelf	0.21	0.03	0.37						0.34	<0.01	0.41		
Atlantic cod	B	GB	0.64	<0.01	0.58										
Haddock	B	GB	0.29	<0.01	0.34	0.30	<0.01	−8.1							
Pollock	B	GOM	0.58	<0.01	0.74*										
White hake	B	GOM	0.30	<0.01	0.54										
Red hake	B	shelf	0.21	0.03	0.47	0.41			<0.01	−37.9					
American plaice	B	GOM	0.79	<0.01	0.74	0.24	0.02	−7.5							
Summer flounder	B	MAB	0.47	<0.01	0.62						0.58	0.02	0.27	<0.01	4.12
Four-spot flounder	B	MAB	0.20	0.04	0.34						0.40	<0.01	0.14		
Yellowtail flounder	B	MAB	0.64	<0.01	0.52						0.50			<0.01	−0.85
Winter flounder	B	MAB	0.45	<0.01	0.44										
Witch flounder	B	GOM	0.52	<0.01	0.94*	0.46			<0.01	−78.8					
Windowpane flounder	B	MAB	0.43	<0.01	0.49										
Atlantic mackerel	S	MAB									0.58	<0.01	0.59		
Butterfish	S	MAB	0.60	<0.01	0.81*						0.21			0.03	1.28
Bluefish	S	MAB													
Black sea bass	B	MAB	0.35	<0.01	0.58	0.21	0.03	−9.7			0.29	<0.01	0.28		
Scup	B	MAB				0.20	0.04	−8.6			0.32	<0.01	0.41		
Redfish	B	GOM	0.71	<0.01	1.12*										
Longhorn sculpin	B	MAB	0.50	<0.01	0.48						0.24	0.02	0.14		
Northern sea robin	B	MAB	0.20	0.04	0.41	0.29	<0.01	−8.8			0.19	0.04	0.21		
Northern sand lance	B	MAB	0.26	0.02	0.37	0.22	0.03	−3.6							
Ocean pout	B	MAB	0.77	<0.01	0.64						0.28	0.01	0.15		
Goosefish	B	shelf	0.24	0.02	0.38	0.50			<0.01	−87.8	0.35	0.01	0.37	0.01	−1.23

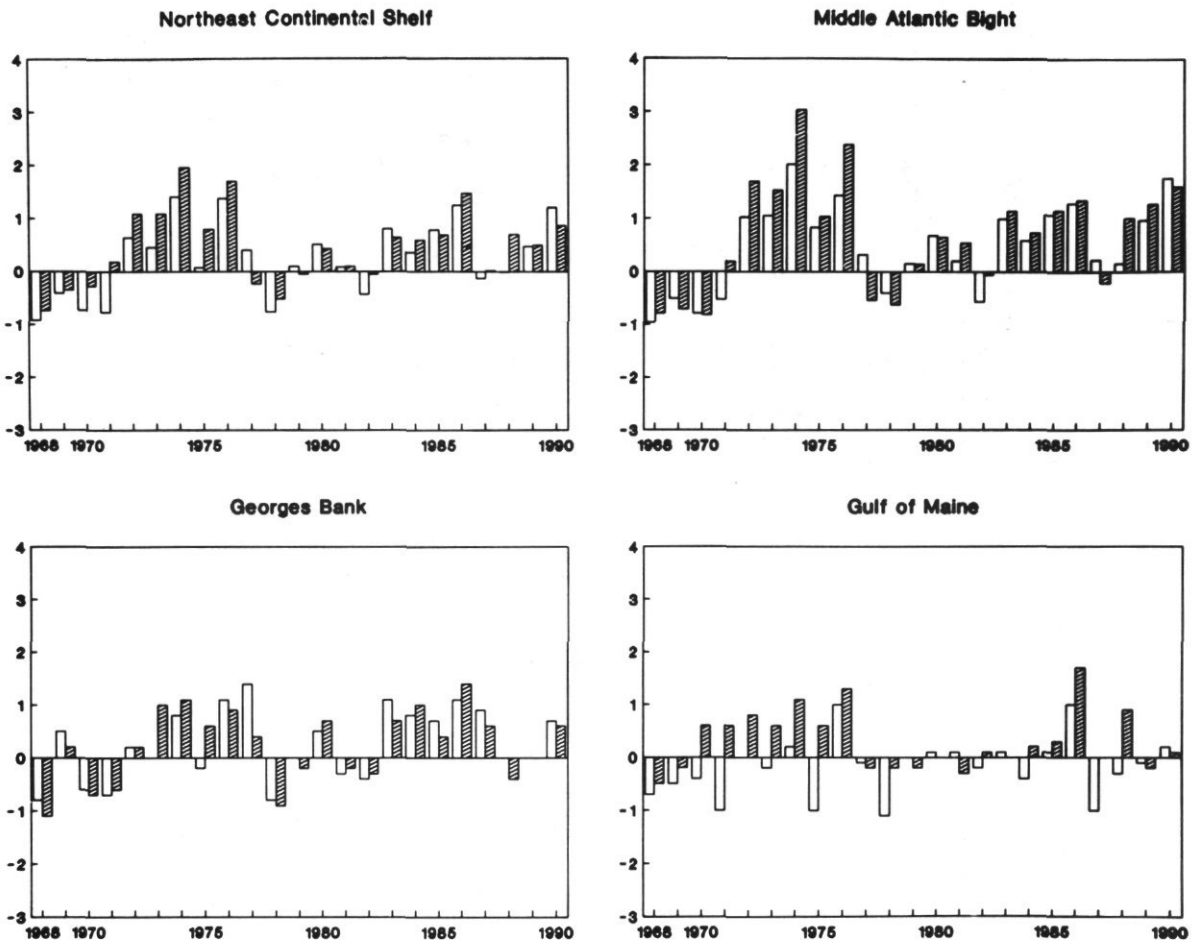

Figure 2. Surface (open bars) and bottom (hatched bars) temperature anomalies (°C) from the spring bottom trawl surveys (1968–1989) for the whole shelf, the Middle Atlantic Bight, the Gulf of Maine, and Georges Bank (adapted from Mountain and Holzwarth, 1990).

increase in the abundance of the elasmobrachs, represented by the skates and spiny dogfish, through the 1980s is believed due to a combination of low fishing mortality of these non-commercial species and an opportunistic response to the decrease in the other fish species.

The principal pelagic stocks (Atlantic herring and Atlantic mackerel) exhibited a substantial recovery in abundance during the 1980s. The increases in both stocks are believed associated with relatively low fishing mortality in recent years (Conservation and Utilization Division, 1991). The increase in herring included a resurgence of herring on Georges Bank, ten years after the fishery there collapsed (Smith and Morse, 1990).

There is no indication that any of the abundance changes illustrated in Figure 3 have been associated with the environmental variability represented in Figure 2. Any consistent environmental influence on the abundance of the fish stocks during the 1980s was greatly overshadowed by the effects of fishing mortality and the resulting changes in the predator–prey interactions between species (e.g. Fogarty *et al.*, 1991).

Distribution: Fish stocks may compensate for interannual changes in shelf water temperatures by changing their location, seeking to maintain some optimum or desirable temperature range. Such compensation can be indicated by comparing the mean temperature of the catch with the average water temperature in the area the species is found. For example, in Figure 4 the mean catch temperature for haddock is plotted against the areal average bottom temperature on Georges Bank. The slope of the regression between the two series is 0.34, meaning that for each degree increase in bottom temperature on Georges Bank the increase in the catch temperature of the haddock was only 0.34°C. The haddock stock compensated for two-thirds of the interannual temperature change by shifting its distribution. When the Bank-wide temperature decreased below about 5.2°C, the haddock stock was found in water warmer than the Bank-wide average. When the Bank

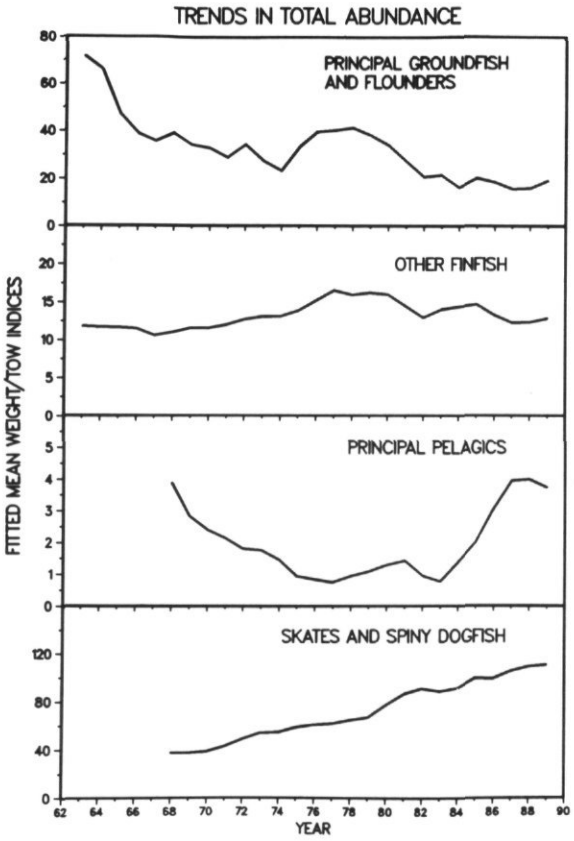

Figure 3. Trends in indexes of aggregate abundance (catch in weight per survey tow) for four species groups, reflecting the major changes in the fishery resources in recent decades (from Conservation and Utilization Division, 1991).

temperature increased above 5.2°C, the stock moved to water cooler than the average.

Twenty-six of the 30 species considered exhibited significant regressions between areal temperature and catch temperature (Table 1). For 22 of these regressions the slope was significantly different from 1, implying some temperature compensation was occurring. The slopes of the regression (Fig. 5) were concentrated between 0.3 and 0.6, meaning that the majority of the species were compensating for about half of the interannual temperature change which occurred.

The average spring surface and bottom temperature distributions (Fig. 6) indicate that a species could compensate for temperature changes by moving either north–south along the shelf or across the shelf between shallow and deeper waters. Using mean catch latitude and mean bottom depth of the catch as indicators of these possible movements, regressions were calculated with areal temperature and catch abundance as independent variables. The results are given in Table 1. Seven species showed a significant relation between areal average temperature and catch depth. In each case increasing areal average water temperature resulted in the species moving to shallower depth, where the waters would be cooler (Fig. 6). Fourteen species had significant regressions between areal average temperature and catch latitude, with increasing temperature resulting in a northward movement in each case, again a movement towards where the water temperatures would be cooler. In three of the fourteen regressions, species abundance was a significant contributing variable, as well. For four species both latitude and depth were related significantly to areal average temperature.

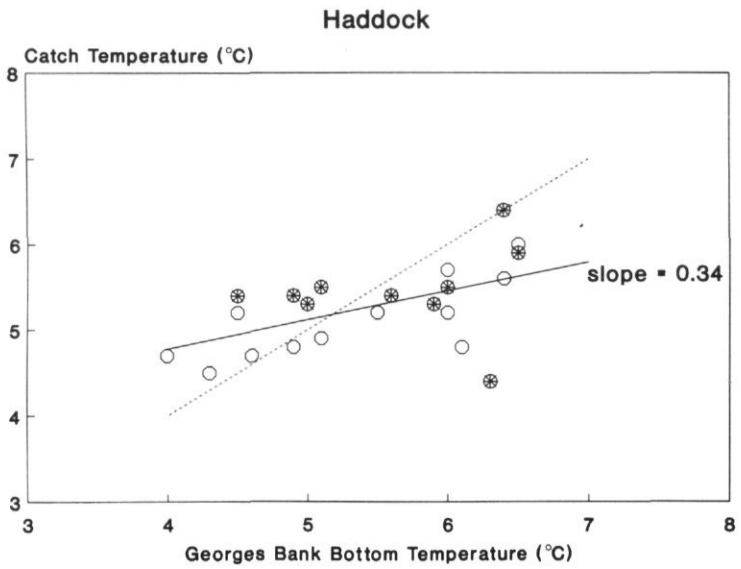

Figure 4. Weighted mean catch temperature for haddock vs. the areal average bottom temperature on Georges Bank from the spring trawl surveys for the period 1968–1989. The points for the 1980s are indicated by filled symbols. The solid line represents the regression of catch temperature on the areal average temperature. The dashed line indicates a slope of 1.

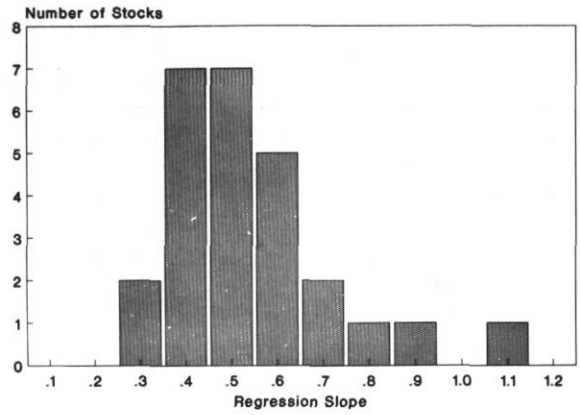

Figure 5. A histogram for the slopes of the significant regressions between catch temperature and areal average temperature for the species listed in Table 1.

Examples of the regressions are shown in Figures 7–9. For mackerel (Fig. 7) the mean catch latitude was related significantly to surface temperature. For each 1°C increase in temperature the stock moved about 0.7° of latitude northward. As would be expected from Figure 2 the points for the 1980s (filled symbols in Fig. 7) are intermediate between the extremes from the earlier years. The haddock stock (Fig. 8) exhibited a relationship between mean catch bottom depth and bottom temperature. For each 1°C increase in average bottom temperature on Georges Bank the stock moved to about 8 m shallower water depth. For spiny dogfish (Fig. 9) both the mean catch bottom depth and latitude were related significantly to the average shelf bottom temperature. Under warmer conditions the species moved both northward and to shallower water. On average for each 1°C increase in temperature the stock moved about

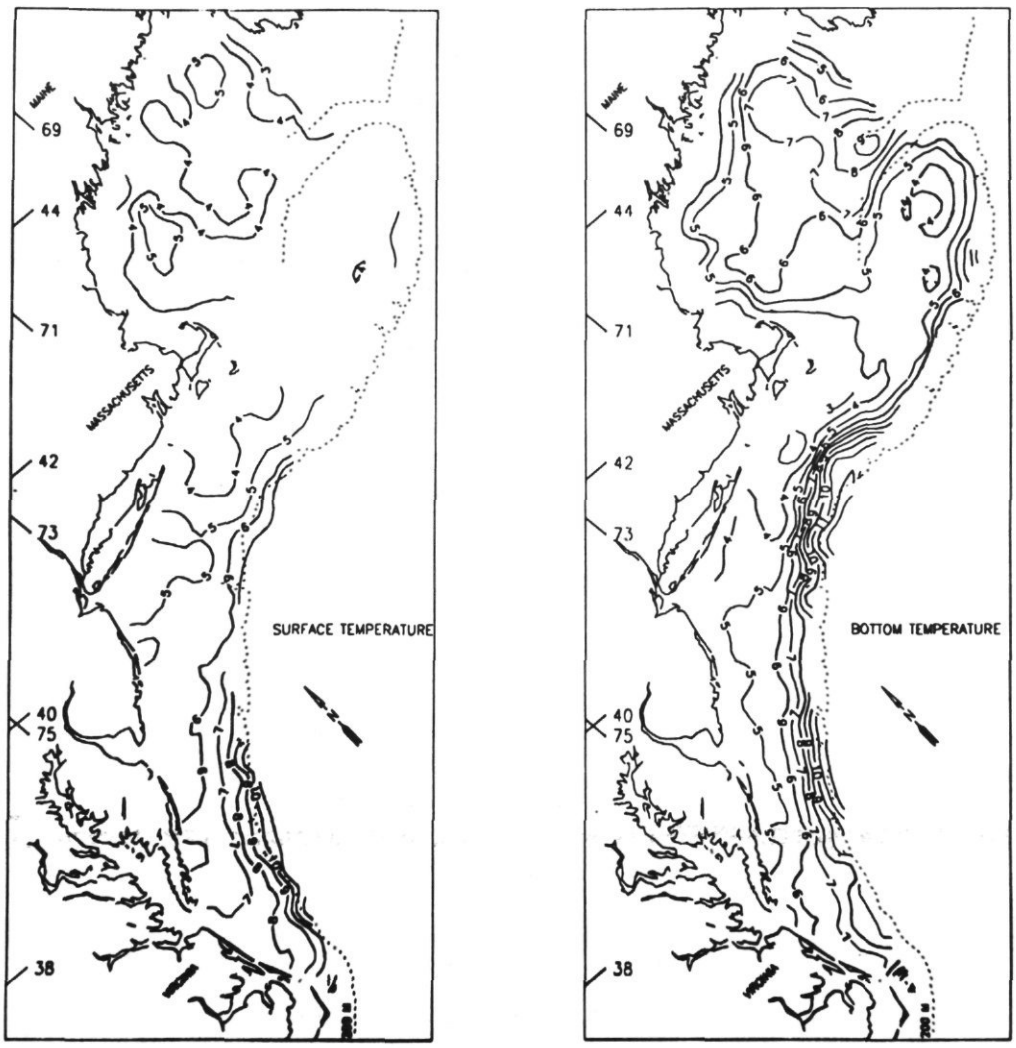

Figure 6. Average surface and bottom temperature distributions (°C) for the northeast shelf of the USA on April 1 (from Mountain and Holzwarth, 1990).

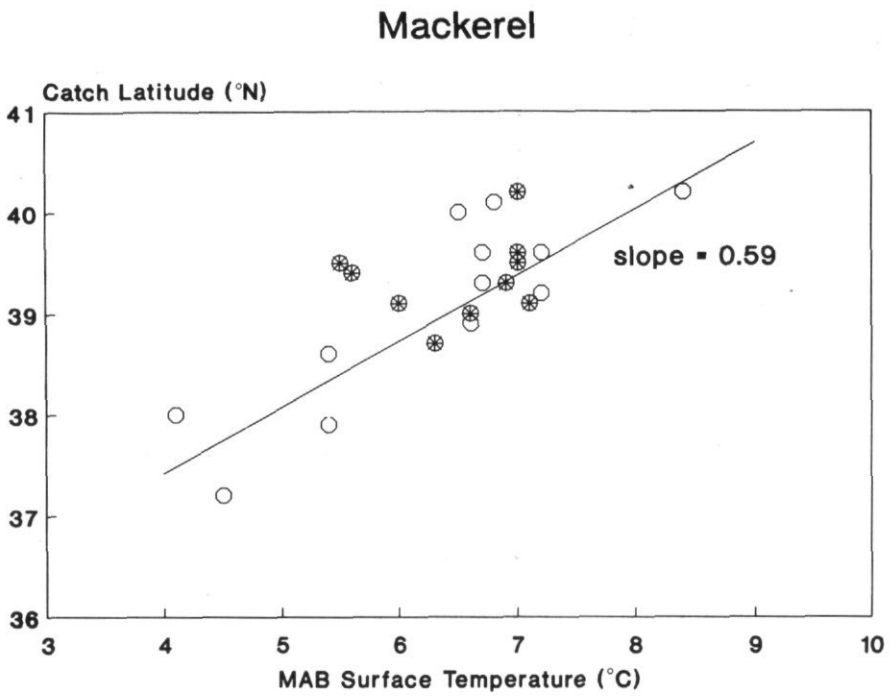

Figure 7. Weighted mean catch latitude for mackerel vs. the areal average surface temperature in the Middle Atlantic Bight from the spring trawl surveys for the period 1968–1989. The points for the 1980s are indicated by filled symbols. The line represents the regression of the catch latitude on the areal average temperature.

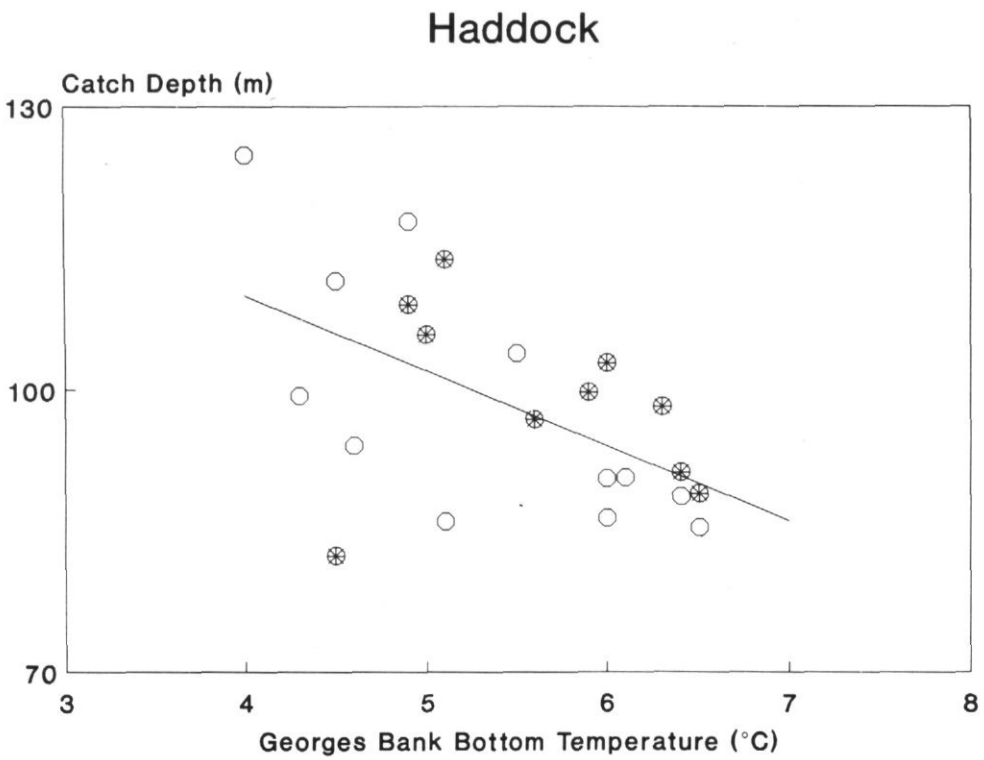

Figure 8. Weighted mean catch depth for haddock vs. the areal average bottom temperature on Georges Bank from the spring trawl surveys for the period 1968–1989. The points for the 1980s are indicated by filled symbols. The solid line represents the regression of the catch depth on the areal average temperature.

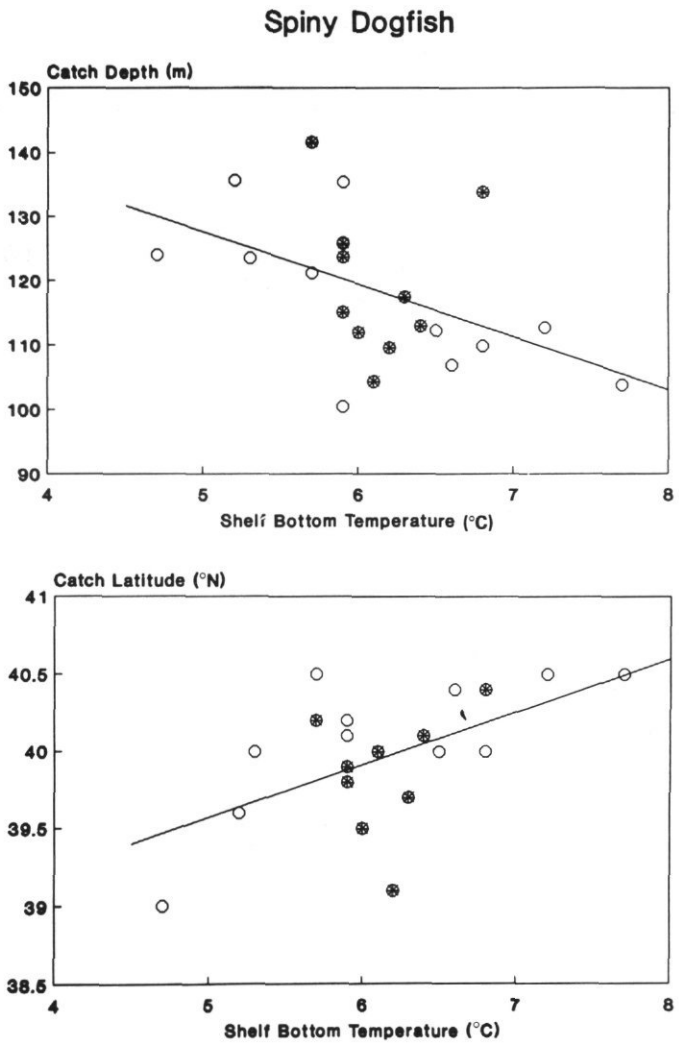

Figure 9. Weighted mean catch depth (top) and catch latitude (bottom) for spiny dogfish vs. the areal average bottom temperature on the northeast shelf from the spring trawl surveys for the period 1968–1989. The points for the 1980s are indicated by filled symbols. The solid lines represent the regression of the catch depth and of the catch latitude on the areal average temperature.

one-third of a degree of latitude northward and about 8 m shallower in depth.

Discussion

Most of the finfish stocks considered compensated in some way to minimize temperature change they experienced when the water temperature on the shelf as a whole changed. The three of the four stocks whose regression slope in Table 1 was not significantly different from one (indicating no significant compensation) were witch flounder, pollock, and redfish. These three stocks are generally found in the deeper waters of the Gulf of Maine, where there is little seasonal temperature change and, perhaps, little tendency for movement in response to temperature changes. The fourth stock was butterfish, which in the spring is generally found near the offshore edge of the MAB where the shelf-wide average temperatures used here are probably not representative.

For species which seasonally migrate, such as mackerel, a relationship between the spring shelf water temperature and catch latitude could represent either a change in the location of the stock over the winter or change in the timing of migration – or both. With the data available one cannot separate these possibilities.

In the haddock catch bottom depth versus bottom temperature regression (Fig. 8) one point (4.5°C, 82 m) was about twice as far from the regression line as any other point. It represents the spring of 1988 and is an example of a different environmental influence on the distribution of a fish stock. In the late spring of 1987 the haddock larval population, which is usually located on

Georges Bank, was observed to be almost entirely in the Middle Atlantic Bight (Polacheck *et al.*, 1992). The westward displacement of the larvae is attributed to an unusually strong westward flow of water along the shelf during 1987 (Polacheck *et al.*, 1992). One-year-old haddock were caught in large numbers in the shallower regions of the MAB during the spring 1988 trawl survey and contributed to a low mean catch depth for haddock in that survey. This situation is unique in the trawl survey records and indicates that circulation events can affect significantly the distribution of a stock in individual years. The recruitment of the 1987 haddock year class did not appear adversely affected by the larval displacement (Polacheck *et al.*, 1992). Dropping the spring 1988 point from the catch bottom depth–temperature regression in Figure 8 increases the R^2 of the regression from 0.30 to 0.48.

Distributional shifts associated with temperature change have implications for recruitment success – and therefore abundance – of the different species and may provide insight into the response of the fisheries ecosystem to anticipated climate change. The magnitude of temperature variability observed since 1968 (Fig. 2) is comparable to the change predicted by models of climate change (Manabe and Stouffer, 1988). Relative distributional shifts could affect the regional competitive balance between species, their predator–prey interactions, as well as the by-catch associated with individual fisheries (Murawski and Mountain, 1990). The net effect on the recruitment of a particular species could be quite complex to determine. The haddock stock on Georges Bank provides an example of this complexity.

A shift by the adult haddock population toward shallower waters with warmer temperature (Fig. 8) potentially would result in a greater portion of the haddock larvae being in the well mixed portion of central Georges Bank. Larval survival is relatively low in the well-mixed waters as compared to the deeper, stratified regions of the bank (Buckley and Lough, 1987). At the same time the larvae would be further from the southern edge of the bank where they can be lost from the system by offbank advection due to wind events or entrainment by warm core, Gulf Stream rings. With a simultaneous northward shift in the mackerel stock (Fig. 7), the arrival of the spring mackerel migration onto Georges Bank and its possible predation on haddock larvae (Michaels, 1991) would occur earlier – which conceivably could increase or decrease the total predation on haddock larvae. The net effect of increasing temperature on the survival of the early life stages of haddock is not clear.

Retrospective correlative analyses, as presented in this report, will need to be combined with the results of specific studies of the processes controlling recruitment to determine the potential effects of climate change on the fish stocks of the Northwest Atlantic.

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Variability of winter ocean conditions and strong year classes of Northeast Pacific groundfish

Anne Babcock Hollowed and Warren S. Wooster

Hollowed, A. B., and Wooster, W. S. 1992. Variability of winter ocean conditions and strong year classes of Northeast Pacific groundfish. – ICES mar. Sci. Symp., 195: 433–444.

This paper compares winter environmental conditions and year-class strength for a set of Northeast Pacific fish stocks. Since 1932, environmental variability has been characterized by alternating warm and cool eras of 6 to 12 years' duration. There appear to be two environmental states: one associated with weak circulation in the Alaska gyre and enhanced upwelling along the west coast of North America the other with strong circulation in the Gulf of Alaska and weak upwelling farther south. The latter situation characterizes the warm eras, which are commonly initiated by an El Niño event. Since 1958, synchronous strong recruitment occurred among a number of groundfish stocks in 1961, 1970, and 1977. Each of these years was warm and exhibited strong circulation in the Alaskan gyre. While the effect of large-scale environmental conditions seems clear, the mechanism by which recruitment is influenced requires further investigation.

Anne Babcock Hollowed: National Marine Fisheries Service, Alaska Fisheries Science Center, Bin C15700, 7600 Sand Point Way NE, Seattle, WA 98115-0070, USA. Warren S. Wooster: School of Marine Affairs HF-05, University of Washington, Seattle, WA 98195, USA.

Introduction

The fishery literature is replete with studies in which recruitment and environmental variables have been correlated, but such relationships seldom have continuing predictive power (Drinkwater and Myers, 1987; Prager and Hoenig, 1989). One reason they break down is that abiotic variabilities and the response of fish stocks to the fluctuations in the environment are usually non-linear. Furthermore, the ecosystem may have more than one equilibrium state, each of which would respond differently to variable environmental conditions (Skud, 1982; Steele and Henderson, 1984; Wooster and Bailey, 1989).

Fishery scientists who study the key processes regulating recruitment of marine fish are interested in extreme recruitment events as well as those contributing to long-term variability in recruitment. Strong year classes are of particular importance because many fisheries depend on them. One might expect such year classes to be uncommon because they usually depend on the conspiracy of several environmental conditions that favor survival at critical stages of development. Poor year classes, on the other hand, might occur when any of several environmental conditions are unfavorable.

High frequency (interannual) and longer term patterns of recruitment of Northeast Pacific marine fishes were described by Hollowed *et al.* (1987). Hollowed (1990) compared years of synchronous extreme (strong or weak) year classes with time series of environmental indices. In this paper, we re-examined the data with a view to identifying a suite of environmental variables that appear to favor the development of synchronous strong year classes of Northeast Pacific groundfish.

Data sources and methods

An extensive time series of oceanographic data was available for the Northeast Pacific region. In this analysis, seven indices were examined: sea level, Northeast Pacific Pressure Index (NEPI), El Niño events, sea surface temperature, upwelling intensity, temperature at 100 m, and estimated dates of the spring transition. These seven indices provide a general indication of oceanic conditions within the Northeast Pacific Ocean.

1. *Sea level*: Sea-level data were collected from coastal stations at La Jolla, San Francisco, and Crescent City, California; Neah Bay, Washington; and Sitka, and Yakutat, Alaska (Fig. 1). These data were obtained from the Pacific Marine Environmental Laboratory (PMEL),

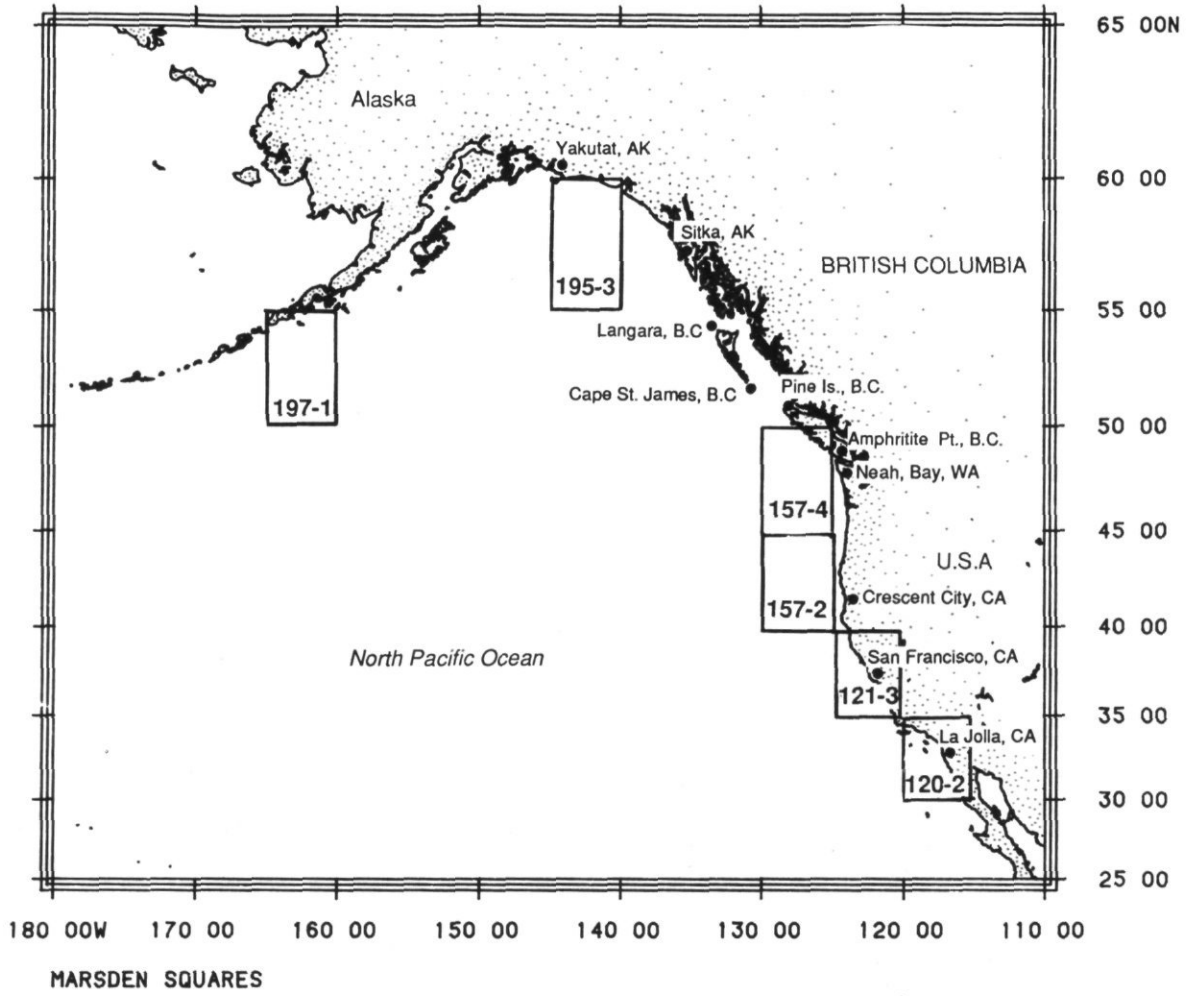


Figure 1. Map of coastal sea level, sea surface temperature stations, and 5° × 5° degree Marsden squares.

Seattle, Washington, and have been corrected for the inverted barometer effect and for isostatic rebound (Cannon *et al.*, 1985).

2. *Atmospheric pressure*: The NEPI time series was included in the analysis to examine the influence of large-scale ocean forcing. The NEPI index represents the seasonal mean pressure difference between 40°N 120°W and 50°N 170°W (Emery and Hamilton, 1985). The published series for 1931–1982 was updated by the Fisheries Oceanography Coordinated investigations (FOCI) group at the PMEL and will be distinguished from NEPI as NEPIFOCI.

3. *El Niño*: Occurrences of El Niño in the Equatorial Pacific Ocean identified by Quinn *et al.* (1987) and Rasmusson and Carpenter (1982) were used in this study.

4. *Sea surface temperature*: Values of monthly mean sea surface temperature by 5° × 5° quadrants of Marsden squares (identified in Fig. 1) were obtained from the Pacific Fishery Environmental Group (PFEG), Monterey, California (Johnson *et al.*, 1976). In some cases the time series were extended by averaging data summarized by 1° × 1° quadrants obtained from PFEG. Sea surface temperature data from 2° × 2° quadrants bordering the Northeast Pacific coast were also obtained from the Comprehensive Ocean-Atmosphere Data Set (COADS) (Woodruff *et al.*, 1987). Additionally, sea surface temperature data from coastal stations in British Columbia, Canada (Amphritite Point, Pine Island, Cape Saint James, and Langara; see Fig. 1) were provided by S. Tabata (Institute of Ocean Sciences, Sidney, BC, Canada).

Missing values of monthly winter (January–March) sea surface temperature were estimated from a regression with months where data were available. In cases when values for two months were available, the estimate was based on the average of the estimated temperature from both regressions. In all cases the regressions were significant ($p < 0.05$).

5. *Upwelling intensity*: Monthly upwelling intensity

was estimated from Bakun indices derived from large-scale geostrophic wind stress (Bakun, 1973; Mason and Bakun, 1986). Monthly values were calculated from synoptic surface atmospheric pressure fields prepared by Fleet Numerical Weather Central (FNWC) arranged on a three-degree mesh length grid. Records were examined from locations centered on the following coordinates: 36°N 122°W, 39°N 125°W, 42°N 125°W, 45°N 125°W, 48°N 125°W, 54°N 134°W, 57°N 137°W, and 60°N 146°W.

6. *Temperature at 100 m*: Estimates of the mean monthly temperature at 100 m were calculated by the PFEG on a three-degree mesh length grid. Records from four locations were examined (36°N 122°W, 39°N 125°W, 42°N 125°W, and 45°N 125°W.

7. *Spring transition*: The spring transition marks the shift from northward monthly mean winter winds to southward summer winds, which results in an intensification of equatorward flow along the coasts of Washington, Oregon, and California. The date of the spring transition was estimated following the procedures described by Strub *et al.* (1987) using time series of daily sea level and daily upwelling indices (Hollowed, 1990).

In this study, we assumed that environmental conditions that influenced survival during the first six months of life determine year-class strength. Because all of the fish stocks considered in this analysis exhibit peak spawning periods during the winter or early spring (Parrish *et al.*, 1981; Garrison and Miller, 1982), the time period for consideration of the environmental indices was limited to the winter months (December, January, February, and March).

Subsets of winter months were selected for each of the environmental variables in the following manner. Monthly indices of sea surface temperature and temperature at 100 m, were highly correlated in the winter. Therefore, a composite index representing the mean of January, February, and March was used for the analysis of sea surface temperature, and temperature at 100 m. Following Emery and Hamilton (1985), the winter NEPIFOCI index was defined as the mean of the December, January, and February pressure indices. Because of the influence of the spring transition from weak southerly winds to persistent northerly winds, two indices of upwelling were considered: the mean value of the January and February upwelling indices and upwelling in March. March upwelling was distinguished from the other months because the spring transition, which typically occurs between March and April, is accompanied by a sharp shift in wind that could bias the mean. The spring transition also has a strong influence on sea level. Therefore, the winter sea-level index was the average of January and February values, excluding those of March.

With the exception of the time series of transition dates, each of the time series of environmental data was standardized by subtracting the long-term mean and dividing by the standard deviation. In the case of the transition dates, extreme years were identified by dividing the time series into quartiles. Years when the transition data was early or late were identified as years in the upper or lower quartile of observed transition dates. The use of quartiles rather than standard deviates in the case of the spring transition dates was necessary because the time series was short and the data were skewed.

Pattern of environmental variability

Comparison of the environmental time series (Figs. 2 and 3) revealed a striking pattern in sea surface temperature. An alternating sequence of cool and warm eras occurred during the period 1932–1988 (earlier data are too sparse for this determination):

1932–1939 cool
1940–1945 warm
1946–1957 cool
1958–1964 warm
1965–1976 cool (interrupted in 1970)
1977–1987 warm (interrupted in 1979)

These periods lasted for 6 to 12 years (average 9.3 years). The three warm eras (40% of the record) appear to have been initiated by specific El Niño events; that of 1940–1945 by the 1939–1941 events, that of 1958–1964 by the 1957–1958 event, and that of 1977–1987 by the 1976 event. The isolated warm year in 1970 may be related to the 1969 event.

Not all El Niño events affect the Northeast Pacific Ocean. For example, there was little evidence of the 1965 and 1972–1973 events, which had profound effects in the Peru Current ecosystem (Valdivia, 1978) (Fig. 3). The massive El Niño of 1982–1983, on the other hand, was clearly evident in the already established warm era.

Most other environmental variables have tendencies that prevail during the eras defined by sea surface temperature; they can be summarized as follows:

	Sea surface temperature	
Environmental variable	Warm	Cool
Temperature at 100 m	High	Low
Sea level	High	Low
NEPIFOCI	High	Low
Transition data	Mixed	Mixed
Upwelling (January–February)	Low	High
Upwelling (March)	Mixed	Mixed

These observations support the view that there are alternating environmental regimes in the eastern North

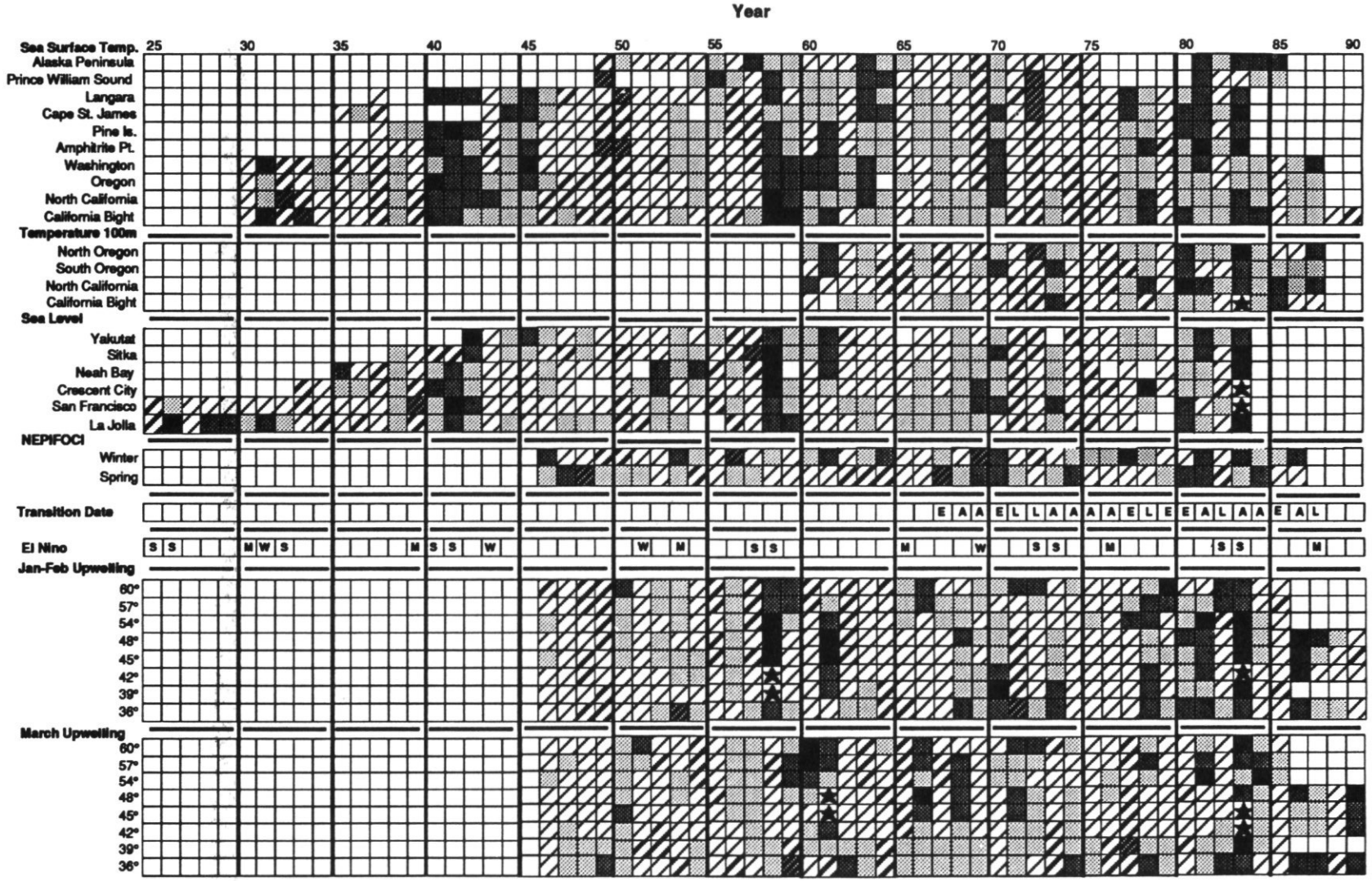

Year
Sea Surface Temp.
25
30
35
40
45
50
55
60
65
70
75
80
85
90
Alaska Peninsula
Prince William Sound
Langara
Cape St. James
Pine Is.
Amphitrite Pt.
Washington
Oregon
North California
California Bight
Temperature 100m
North Oregon
South Oregon
North California
California Bight
Sea Level
Yakutat
Sitka
Neah Bay
Crescent City
San Francisco
La Jolla
NEPIFOCI
Winter
Spring
Transition Date
E A A E L L A A A A E L E E A L A A E A L
El Nino
S S M W S M S S W W M S S M W S S M S S M
Jan-Feb Upwelling
60°
57°
54°
48°
45°
42°
39°
36°
March Upwelling
60°
57°
54°
48°
45°
42°
39°
36°

Figure 2. Standardized environmental data summary. Shading codes for SST, T_{100}, sea level, NEPIFOCI, and upwelling are shown in the key. El Niño: S, strong; M, medium; W, weak. Spring Transition: E, early; A, average; L, late.

Pacific Ocean, which we identify as Type A and Type B (Fig. 4). This view was advanced by Chelton and Davis (1982) who examined sea-level data and proposed that the Alaska and California Currents fluctuate out of phase, with enhanced circulation in the Alaska gyre being accompanied by reduced southward transport in the California Current and vice versa. Chelton (1984) also noted that the transition between oceanic regions may vary considerably depending on the northern or southern extent of the west wind drift and the subsequent bifurcation of the Alaska and California Currents.

In our interpretation of the data, Type-A conditions are characterized by weak circulation in the Alaska gyre (indicated by low NEPIFOCI values), with strong upwelling farther south leading to low temperatures and low sea level. In the Type-B state, a well-developed Pacific high-pressure cell and an intense Aleutian low-pressure cell cause NEPIFOCI to be high; strong circulation in the Gulf of Alaska is accompanied by relatively high sea surface temperatures and sea level and deep thermoclines farther south.

Warm eras and isolated El Niño events tend to be associated with Type-B circulation, cool eras with Type-A circulation. While the warm eras appear to be initiated by El Niño events (but not by all El Niño events), it is not obvious how their duration or termination (transition to cool eras) is determined.

Other studies of low-frequency environmental change in the region are compatible with this scheme. For example, McLain (1984) describes the transition in winter ocean conditions that occurred in 1976. In the early 1970s, winter conditions were characterized by a well-developed ridge pattern of upper air circulation over the central North Pacific Ocean. After the transition the ridge shifted to a position over western North America. Winds along the coast then became more southerly, and onshore transport piled water up against the coast depressing the thermocline. Since 1976, the values of the Southern Oscillation Index have been continually low and weak El Niño conditions persisted with above normal sea surface temperatures, sea levels, and rainfall in winter.

Emery and Hamilton (1985) also found two characteristic distributions of seasonal mean sea level pressure, with strong circulation associated with an intensified Aleutian low and weak circulation associated with its westward displacement. They found that strong sea level pressure patterns (as in our Type-B state) often coincide with El Niño events.

Niebauer (1988) traced the Southern Oscillation relationship farther north to the eastern Bering Sea. He showed that warming in the Bering Sea tends to follow El Niño events. The linkage is presumably through the atmosphere because ocean exchange through the eastern Aleutian Island passes is limited. Niebauer showed that the winter Aleutian low was intensified and farther east than normal during warm (El Niño) events.

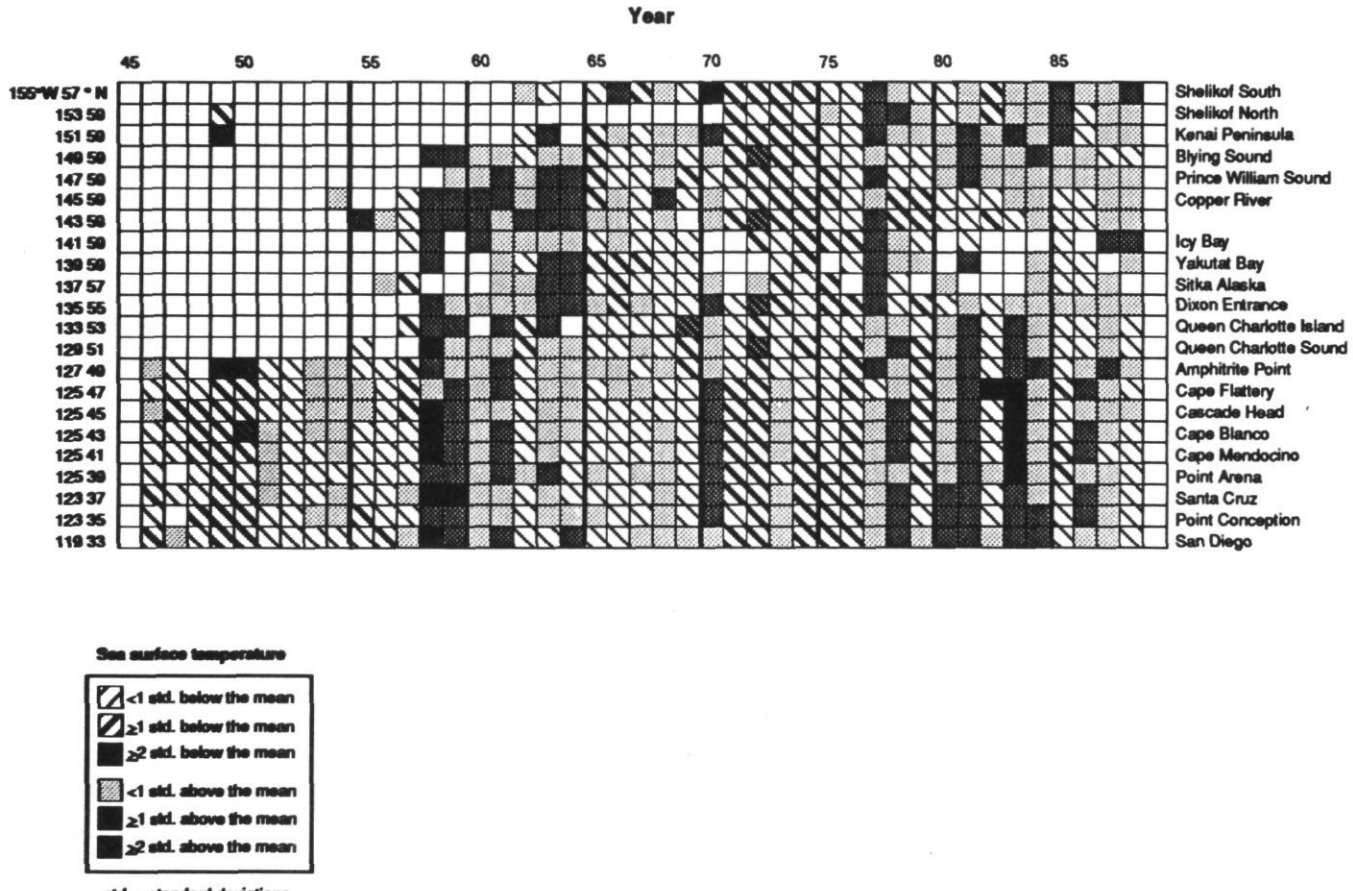
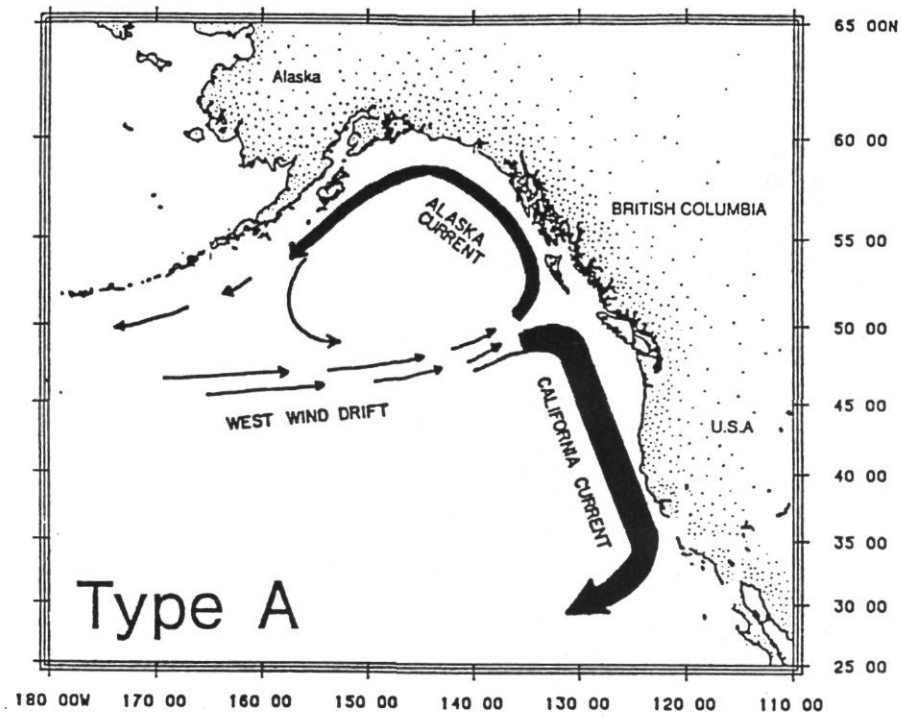


Figure 3. Summary of standardized sea surface temperature from 2° × 2° squares. Shading codes are shown in the key.



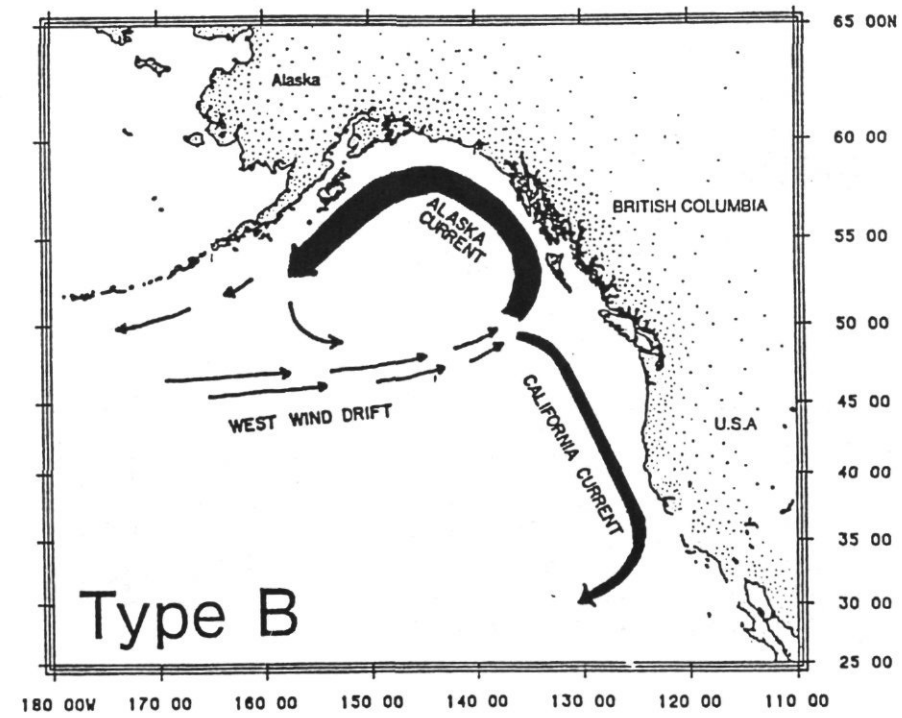


Figure 4. Diagram of circulation patterns associated with Type-A and Type-B ocean conditions.

Roden (1989) analyzed data from the west coast of North America and identified three recent temperature periods (cold, 1948–1956; warm, 1940–1945, and 1977–1984) as well as two earlier periods (cold, 1857–1863; warm, 1864–1870). Low frequency fluctuations in sea surface temperatures at and north of 53°N in the Northeast Pacific Ocean were studied by Royer (1989). Royer concluded that these fluctuations were not forced by wind but may instead be associated with changes in solar activity.

Tabata's (1989) examination of temperature data at 10 decibars at Ocean Station P can also be related to our study of surface variability. Our interpretation of his data suggests a warm era from 1957 to 1963, a cool era from 1964 to 1977, and a warm era from 1978 to at least 1983. Although these periods were determined from open ocean data, they differ from our more coastal determination by only one year.

A more striking contrast appears below 100 m, where the recent warm era apparently began in 1972 and extended only through 1981. Tabata points out that the subsurface anomalies are related to vertical displacement of isopycnals which are deeper when temperatures are high. Ocean Station P (at 50°N, 145°W) is located within the eastward Subarctic Current, and the deepening of isopycnals may indicate intensification of that flow or northward displacement of the gyre.

Ocean conditions and strong groundfish year classes

Environmental conditions were so different from era to era that they would be expected to have different consequences for the success of recruitment of fish stocks. These differences are likely to have profound influences on recruitment of marine fish stocks.

To explore this possibility, time series of recruitment indices for 15 groundfish stocks were examined for evidence of synchronous strong year classes (Table 1). The occurrence of synchronous strong year classes across a large geographic range is seen as strong evidence of the influence of favorable environmental conditions on fish stocks (Shepherd *et al.*, 1984).

Extreme year classes were identified as described in Hollowed *et al.* (1987). High-frequency variability in recruitment was estimated from the residual of observed recruitment (X) from the five-year running median (RM) represented as a proportion of the running median (PRS):

$$PRS(t) = [X(t) - RM(t)]/RM(t).$$

Thus, the residuals from the RM represented recruitment success or failure relative to adjacent years and was most useful in detecting short-term variations in recruitment. One drawback to this approach is the loss of the first and last two years of the time series. Strong year classes were defined as those instances when the PRS was in the uppermost quartile. Synchronous strong year classes were defined as years when 50% or more of the groundfish stocks had strong year classes.

Synchronous strong year classes were identified in 1961, 1970, 1977, and 1984 (Figs. 5 and 6). The strong year classes observed in 1962 were considered an artifact of ageing errors (Hollowed, 1990). Time series of recruitment were available for only four stocks in 1984. Therefore, classification of this year as a year of synchrony may be premature. The time series of recruitment for the west coast sablefish stock that appeared in Hollowed (1990) was not included in this analysis because of uncertainty in the quality of the recruitment estimates (R. Methot, Alaska Fisheries Science Center, Seattle, WA, pers. comm., August 1991).

Table 1. Summary of quantitative data sources.

Stock	Scientific name	Spawning location	Source
Pacific hake	*Merluccius productus*	CA	Dorn *et al.* (1991)
Dover sole	*Microstomus pacificus*	WA-OR	Demory *et al.* (1984)
POP (Vancouver)	*Sebastes alutus*	WA-OR	Ito *et al.* (1986)
POP (Columbia)	*Sebastes alutus*	WA-OR	Ito *et al.* (1986)
Petrale sole	*Eopsetta jordani*	WA-OR	Demory (1984)
English sole	*Parophrys vetulus*	WA-OR	Golden *et al.* (1986)
Widow rockfish	*Sebastes entomelas*	CA-WA	Hightower and Lenarz (1989)
Pacific halibut	*Hippoglossus stenolepis*	CAN-GOA	I.N.P.H.C. (1991)
Rock sole	*Lepidopsetta bilineata*	CAN-HS	Fargo (1989)
English sole	*Parophrys vetulus*	CAN	Fargo (1990)
Petrale sole	*Eopsetta jordani*	CAN	Ketchen and Forrester (1966)
Pacific cod	*Gadus macrocephalus*	CAN-WVI	Tyler and Foucher (1990)
Pacific cod	*Gadus macrocephalus*	CAN-HS	Tyler and Foucher (1900)
POP	*Sebastes alutus*	CAN-QC	Archibald *et al.* (1983)
Walleye pollock	*Theragra chalcogramma*	GOA	Hollowed and Megrey (1989)

CA: California, WA–OR: Washington–Oregon; CAN: Canada; CAN–HS: Hecate Strait; CAN–WVI: West coast of Vancouver Island, Canada; CAN–QC: Queen Charlotte; GOA: Gulf of Alaska.

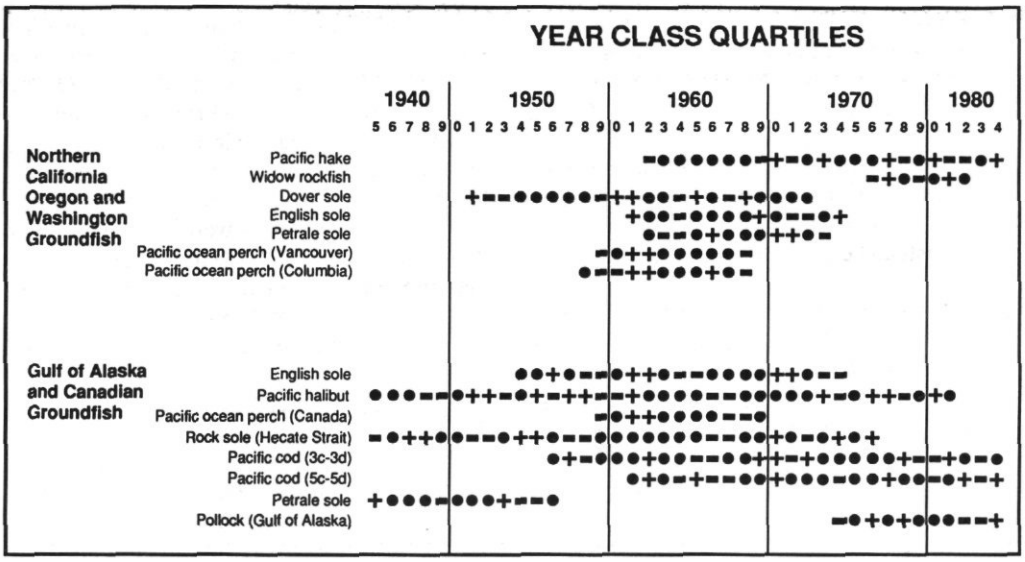


Figure 5. Summary of strong and weak year classes for Northeast Pacific marine fish stocks. The plus and minus signs indicate extreme year classes, and the solid dots represent an average year class.

Is there a unique set of environmental conditions associated with recruitment success? The answer can be found in Figures 2 and 3. Each of the years of synchronous strong year classes occurred during a warm era. Furthermore, most took place on the first occasion when winter NEPIFOCI values were more than one standard deviation above the mean, either at the start of the era (1977) or after the era was under way (1961). These conditions also apply in the isolated warm year 1970. In the terminology used earlier, Type-B circulation prevailed during each year of synchronous strong groundfish year classes. Three of the years of synchronous strong year classes were associated with El Niño events (i.e. 1970, 1977, 1984).

Before the warm era initiated in 1958, synchronous strong year classes of Northeast Pacific groundfish were identified in 1951. This year occurred during a cool regime when environmental conditions were not un-

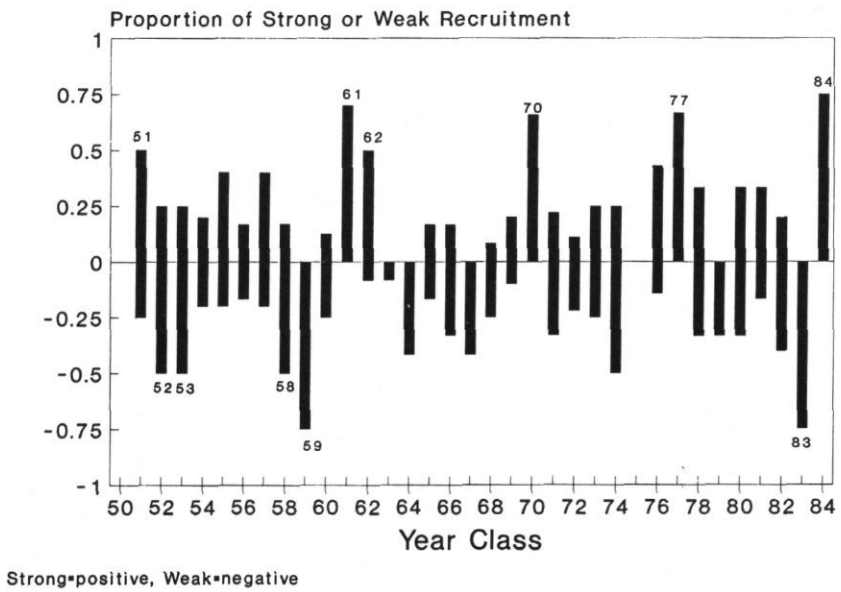


Figure 6. Summary of the percentage of Northeast Pacific groundfish stocks that had extreme year classes. Years when 50% or more of the stocks were extreme are indicated.

usual. These early data include only a few species, but if they generally represent groundfish stocks, a regime shift may have occurred in 1958 under the influence of the strong El Niño event of 1957–1958. The observed set of environmental conditions favoring strong groundfish year classes after 1958 may not have favored recruitment prior to 1958. For example, 1958 might have been expected to produce strong year classes of Northeast Pacific groundfish because sea surface temperature conditions were extremely warm, sea level was high throughout the study region, and winter upwelling was weak. These conditions are similar to those accompanying synchronous strong recruitment observed in 1961, 1970, and 1977, except that unusually high winter NEPIFOCI values were not observed; most Pacific groundfish species exhibited weak recruitment in 1958 (Fig. 6).

These observations support the contention that a set of environmental conditions influence year-class strength of marine fish stocks. While sea surface temperature provides a convenient index for classification purposes, the alternating eras differ in more profound ways because of changes in the patten and strength of circulation in the North Pacific Ocean.

Our finding of a link between synchronous strong groundfish year classes and ocean conditions is reminiscent of Sætersdal and Loeng (1987). In this paper, the authors noted that most strong and medium year classes of Northeast Arctic cod were associated with positive temperature anomalies in the early part of a warm period or they occurred immediately prior to a shift to a warmer ocean regime. They hypothesized that reproduction of Arctic cod is adjusted to variations in the feeding area caused by climatic fluctuations, which occur on a large spatial and temporal scale.

In the present study, we did not advance hypotheses regarding possible mechanisms whereby environmental conditions act to influence the success of recruitment or marine survival. Ocean conditions associated with a particular circulation pattern influence marine fish by altering advection, turbulence, or physiological processes. These processes may influence the behavior of fish, as well as rates of starvation, predation, and infestation. Identification of patterns in relationships between environment and recruitment success may provide clues to the mechanisms linking environment and recruitment.

Postscript

Our paper focuses on events in the Northeast Pacific Ocean and is somewhat out of place in a symposium designed to describe hydrographic variability in the ICES area, 1980–1989. However, throughout the symposium it became clear that there are several important similarities between the two oceanic systems. In the first session, several papers presented time series that demonstrated alternating periods of warm and cool temperature (Ellett and Blindheim, Loeng *et al.*, Svendsen and Magnusson, and Rodionov and Krovnin, all in these proceedings). Loeng *et al.* draw a conclusion similar to ours that alternating temperature eras are related to patterns of large-scale circulation. It is particularly interesting to note that in recent years the temperature time series from the Kola section shows a pattern of variability that is the opposite of that observed in the Northeast Pacific Ocean (Rodionov and Krovnin, these proceedings):

Kola section	Northeast Pacific
Warm (1970–1976)	Cold (1965–1976)
Cold (1977–1982)	Warm (1977–1987)

A warm era with above average ice indices was initiated in 1983 (Ellett and Blindheim, these proceedings, Fig. 9). Strong year classes of several fish stocks, both demersal and pelagic, were noted throughout the northeast Atlantic Ocean in 1983 (Jakobsson, these proceedings). This finding is consistent with our findings and those of Sætersdal and Loeng (1987) that indicated synchronous strong year classes of marine fish stocks are often associated with the initiation of warm eras. Such similarities between distant ocean systems demonstrate the importance of continuing collaborative research efforts in the future.

Acknowledgments

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Analysis of trends in hydrochemical parameters in the Western Baltic in the 1980s

Hans Peter Hansen

Hansen, H. P. 1992. Analysis of trends in hydrochemical parameters in the Western Baltic in the 1980s. – ICES mar. Sci. Symp., 195: 445–454.

Trends in the hydrographic properties and nutrient concentrations of the Western Baltic from 1980 to 1990 are identified from national and HELCOM data (GESPA) and compared with those of the preceding decades. Found to be generally weak, these trends are analysed and discussed as integrated effects of the biological and chemical sink capacities of the ecosystem, natural effects (meteorological and hydrographic changes and water exchange), and anthropogenic factors. Measurements in the Kiel Bight are used to study modifications of annual cycles due to climatic fluctuations. It is demonstrated that the Kiel Bight area is representative of a considerable part of the Baltic Sea. The generation of artificial trend signals is discussed as a consequence of non-matching scales of processes and sampling frequencies and of changes in annual cycles, and effects of non-random properties of the data used for statistical treatment. The results of this discussion are used to evaluate the trends observed during the past decades. The congruence of the observed nutrient trends in the Kiel Bight and the descriptive and prognostic results of the Wulff and Stigebrandt budget model for the Baltic Sea and some subareas are examined.

Hans P. Hansen: Institute for Marine Research, Düsternbrooker Weg 20, D–2300 Kiel 1, Germany.

Introduction

The particular situation of the Baltic Sea has drawn scientific as well as public interest to temporal changes which might reflect anthropogenic disturbance of the environment. Relevant time scales range from years to decades or even centuries. The identification of systematic changes within these scales and their causal relation to natural and/or anthropogenic processes are the basis for possible measures to improve the environmental state of the Baltic Sea. In the course of this presentation I use the term "trend" for systematic changes of at least five years, while "long-term trends" are calculated for periods of several decades.

The area

The actual hydrobiological property of sea water is the consequence of numerous processes involving a wide variety of time scales and resulting in a considerable spatial variability (patchiness). For practical reasons the collection of time-series data has to be restricted to a limited number of representative stations, where "representative" means: situated in a horizontally homogeneous water body of uniform vertical variability.

As a matter of fact, the definition of a homogeneous location always requires the compromise of ignoring spatial variability within the limits set by the actual scientific problem. However, the basic characteristics and the dominating processes are identical for the Kattegat, the Belt Sea, and the Baltic Proper (Fig. 1).

The haline stratification separates deeper areas into surface layer and bottom water and reduces vertical exchange. A particular feature of the Baltic Sea, especially the western areas, is that, on average, the low saline surface water is outflow from the Baltic and the deep water of higher salinity is inflow of North Sea or Atlantic water. During periods of reduced or missing inflow the deep water is kept under stagnant conditions. A permanent input of material from land and atmosphere is counteracted by internal sinks, sedimentation, exchange with adjacent areas (or, in the case of nitrogen,

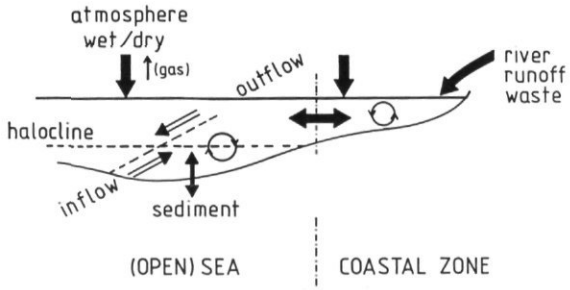

Figure 1. Exchange processes controlling the nutrient concentrations in the Western Baltic.

with the atmosphere) and, finally, exchange with the Atlantic.

The input from land has to pass a coastal area with specific processes which may act as filter for the input material.

The basic hypothesis is that net sink capacities of the Baltic Sea are relatively constant, allowing changes (increases) in inputs to change (increase) the concentrations in the water correspondingly. The inversion of this hypothesis states that trends and long-term changes in concentrations would reflect changes in inputs.

Long-term trends

The dominating hydrobiological parameters are phosphorus and nitrogen compounds (phosphate and organic phosphorus, nitrate, nitrite, ammonia, and organic nitrogen), oxygen, temperature and salinity. With respect to the reliability of analytical results, trend investigations may consider data back to at least 1960. Though some parameters are sufficiently reliable before that date, data are too scarce for systematic trend statistics. Related to the problem of eutrophication, primary production would be an important parameter; however, the quality of data older than some 15 years does not seem to justify trend analysis.

One of the main problems in the compilation of data for long-term trend analysis is the fact that only very few stations are serviced sufficiently frequently to produce reliable data bases. "Good" stations are serviced once a month, some national stations even more often; however, national stations are normally coastal stations and less representative of larger areas.

The assiduous work of the experts for the second periodic HELCOM assessment of the Baltic Sea has provided us with a most comprehensive compilation of trends to which we have added a set of data from the Kiel Bight (Table 1).

The expressed annual cycles in all of the hydrobiological parameters suggest that for trend analyses only a selected proportion of the year's data is useful. The common decision is to take winter data (Jan+Feb), where biological activity and inhomogeneity are low. Because the plankton growth is based on the supply of nutrients in the photic layer, the surface values seem to be the most significant data.

Unfortunately, the selection criteria of the data in Table 1 differ a great deal. Trends are calculated for various periods from 5 to 25 years, some for surface waters, some for deep waters or the entire water column. However, despite the marked variability in results and regional differences, we may derive a reliable description of the long-term trend development in the western areas of the Baltic Sea.

The observed long-term trends in the winter surface layer are very similar for all areas of the Western Baltic. Inorganic nitrogen and phosphate show positive trends if calculated back to 1965 or 1960. The trends decrease or even invert the more the calculation period is reduced toward the last decade. Phosphate concentrations have roughly doubled since the 1960s (0.3 μmol dm^{-3} to 0.6 μmol dm^{-3}) and inorganic nitrogen has reached nearly threefold concentrations (4.5 μmol dm^{-3} to >13 μmol dm^{-3}). Silicate has significantly decreased, in some areas to less than half the 1960s concentration.

The trends show systematic tendencies during the last three decades. Trends calculated back to 1970 or before are significant and strong (positive for inorganic nitrogen and phosphate, negative for silicate). Ranging the calculation from today to about 10 years back results in weak trends (still mostly positive for inorganic nitrogen, but slightly negative for phosphate). This is much more obvious in the graphs and seems to be a general feature for the Western Baltic and the Baltic Proper. Figure 2 shows inorganic nitrogen and phosphate concentrations in the winter surface water of the Kiel Bight back to 1960 (a–b) and the corresponding data from Gotland Sea stations (c–d, Nehring *et al.*, 1990). The observed changes are nearly congruent.

Figure 3 displays the atmospheric wet deposition over central Sweden which has increased at a rate of 0.73 mmol $m^{-2}yr^{-1}$ from about 10 to nearly 40 mmol $m^{-2}yr^{-1}$ during the last 35 years. An interesting feature of the graph is that splitting the trend into three periods shows the same feature as for the nutrients in winter surface waters.

Until 1980 the calculated loads from land and atmosphere and the concentrations of inorganic nitrogen and phosphate were closely correlated. It seemed justified to relate the strong increase of phosphorus and nitrogen during 1970 to 1980 to the increasing anthropogenic inputs.

Eutrophication has increased since 1960. This is supported by observations of elevated primary productivity and by increasing contents of organic matter in sediments. Unfortunately most of the primary productivity data are not measurements of production rates but production capacities. The calculation of production rates from capacity measurements requires a number of assumptions which reduce the reliability of the results.

Table 1. Annual accumulation rates [μmol dm^{-3}yr^{-1}] (Nehring *et al.*, 1990; Trzosinska *et al.*, 1990; BIOMON, 1990).

Ar	Sa	Years	Phos.	ΣN	Sili.	TOT N	TOT P	Comment
R	5	74–88		0.42				surf., wi.
R	5	74–88 83–88		0.35 0.74[1] 0.36[2]	−0.55		0.032	waterc. wi. [1] R6 = 2.07 [2] exclus. R6
R6	1	65–89	0.017	0.21[3]				surf., wi. [3] nitrate
R6	1	65–89	0.013	0.1[3]				deep, yr
N	4	79–89	−0.01	0.09	−0.24	1.11	−0.07	surf., wi.
N	4	60–89	0.023	0.28				surf., wi.
N	4	79–89	−0.02	−0.25	−0.78	0.70	−0.09	deep, wi.
N	4	79–89	0.01	−0.04	0.20	0.01	0.01	surf., yr
N	4	79–89	0.03	0.03	0.84	−0.06	0.01	deep, yr
N	2	74–88		0.35				surf., wi.
N	2	74–88		0.14				deep, wi.
N	2	74–88 83–88		0.46 0.11	−0.49		0.04	waterc., wi.
M	4	70–89	0.014	0.13[3]				surf., wi.
M	4	80–89		−0.26[3]				[3] nitrate
M	1	74–88		0.25				surf., wi.
M	1	74–88		0.08				deep, wi.
M	1	74–88 83–88		0.26 0.54	−0.57 −0.27		0.27	waterc., wi

Ar = area/station code, Sa = numbers of pooled stations, surf. = surface, wi. = winter, waterc. = entire water column.

Trends in the 1980s

The "easy" times are gone for the evaluation and interpretation of trends. As shown above, the increases of N and P in winter surface waters in the 1970s due to increasing loads were high enough to override most other processes. So the quantification may be debatable, but the trends were statistically significant and the interpretations and the conclusions acceptable. Speaking in physical terms, the signal we were looking for was strong compared to the noise level. Maybe this is the reason why most experts preparing the 2nd Periodic HELCOM Assessment of the Baltic Sea preferred to calculate trends back to the 1970s or 1960s.

As mentioned above, most international stations are serviced once a month or even less. This means that a trend analysis of for example winter surface waters for a period of 10 years is based on less than 30 values.

Figure 4a represents the phosphate daily means at the Fladen station (after Engström and Fonselius, 1989). The accumulation rate of 0.017 μmol dm^{-3}yr^{-1} is statistically significant for the 22 years' period.

Calculating the accumulation rates for 10-year periods (Fig. 4b), however, gives a good impression of the random nature of trend figures for 10 years at a single position. More realistic 10-year trends may be calculated if data from "similar" stations are pooled, where "similar" means that the dominating processes controlling the concentrations are identical. If the absolute values differ too much they may be normalized. Such trends are given in Tables 1 and 2 for the areas N and M (defined in HELCOM BMP-Guidelines). They show insignificant accumulation (or even decrease) of phosphate, less accumulation of inorganic nitrogen in the Kiel Bight, and a considerable decrease (–0.26 μmol dm^{-3}yr^{-1}) in the Bay of Mecklenburg. The negative trend in silicate is continued.

The literature offers us only few descriptions of changes within the 1980s but there are numerous reports comparing findings in the 1980s to the 1970s, e.g.

- Increase of N and P in surface and bottom waters.
- Increase of N-load from the atmosphere, especially an increase of ammonia.
- Increased primary production (25–30%).
- Decrease in oxygen concentrations.
- Spreading of low-oxygen or H_2S in bottom layers.
- Increased number of dead fish and invertebrates in benthic catches.

These observations may represent delayed consequences of the drastic changes during the 1970s or are related to processes within the 1980s which are not reflected in the long-term trends. For a more detailed analysis of processes within the short period of the last 10

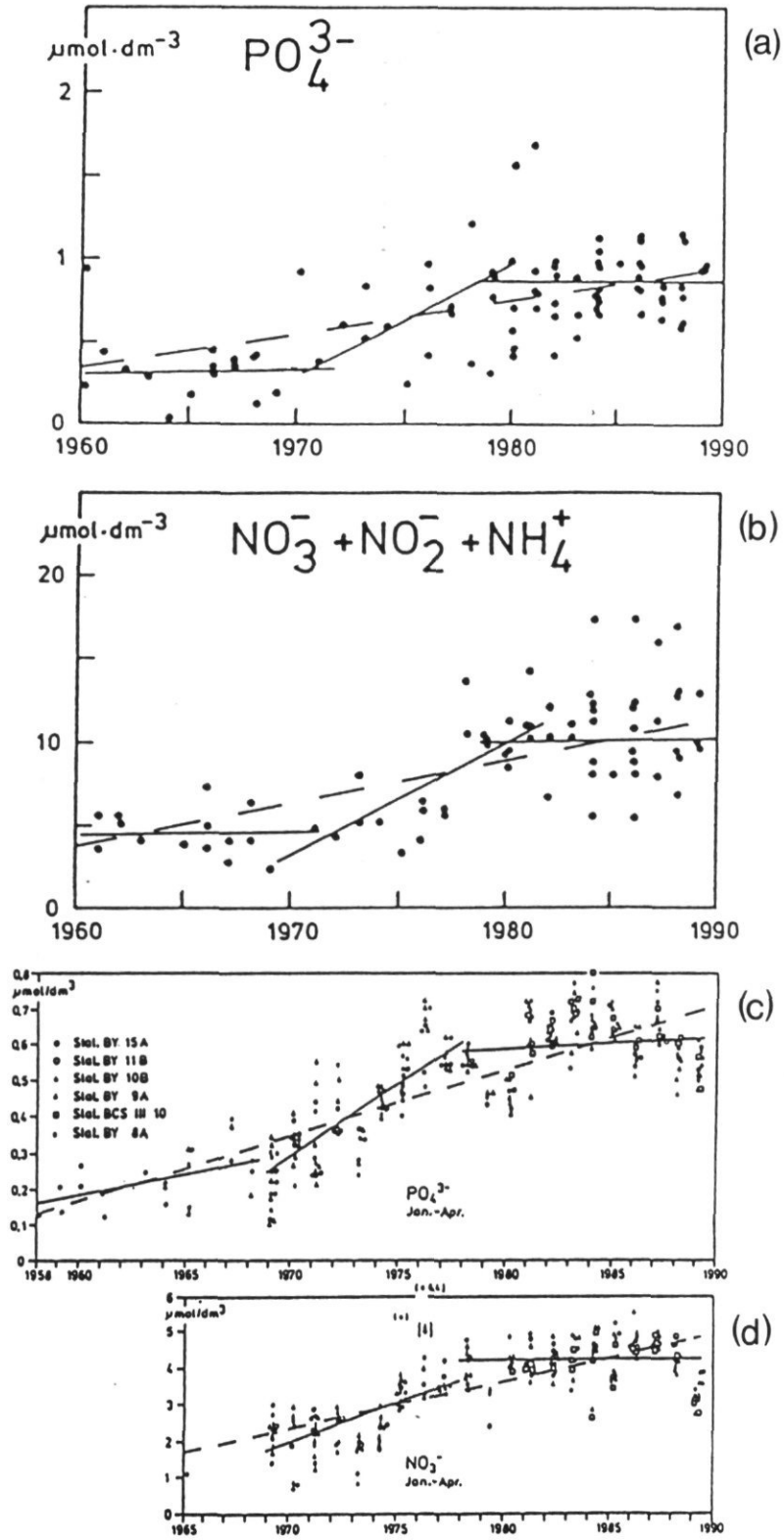

Figure 2. Long-term trends of nutrients (Nehring *et al.*, 1990, original figures adjusted). (a) PO_4^{3-} and (b) ΣN, Kiel Bight, January + February. (c) PO_4^{3-} and (d) NO_3^{-3}, Gotland Sea, January–April.

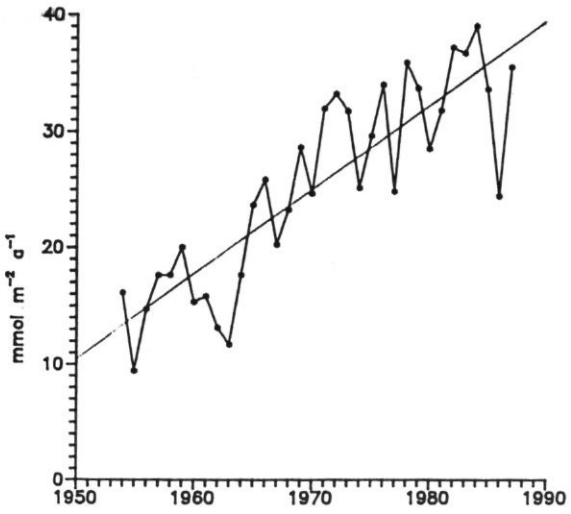

Figure 3. Atmospheric wet deposition of NO_3^- and NH_4^+ over Central Sweden (after SEPA, 1990). Accumulation rate: 0.73 mmol $m^{-2}a^{-1}$.

Table 2. Trends of hydrological parameters in the Kiel Bight from 1979 to 1989 (Area N, four stations pooled).

	Jan+Feb		Jan–Dec	
	surf	bott	surf	bott
Temperature [°C]	0.25	0.19	0.13	0.23
Salinity [PSU]	0.25	0.17	−0.06	−0.05
tot N [μmol dm^{-3}]	1.11	0.70	0.01	−0.06
tot P [μmol dm^{-3}]	−0.07	−0.09	0.01	0.01
Oxygen [$cm^{-3}dm^{-3}$]	-0.06	−0.07	−0.01	−0.10
[1] Aug-Sep			0.01[1]	−0.08[1]

years we have to concentrate on data from an example area with small spatial variability and high sampling frequency. For reasons of convenience (availability) we used data from stations in the Kiel Bight.

Results from the Kiel Bight

The Kiel Bight (Fig. 5) may be regarded as a model for the Baltic Sea. Most of the deeper areas (>25 m) show a permanent halocline. From the inflow to the Baltic Sea through the Great Belt and the Fehmarn Belt, water of higher salinity is pushed through a system of channels and small basins. During periods of missing or weak inflows some of the basins turn to stagnant conditions with sometimes anoxic bottom conditions, e.g. the national station Boknis Eck from which time series back to 1960 and many detailed studies are available. This area is well covered by both national and international BMP stations. All discharges from land and meteorological parameters are monitored and well documented.

The changes in nutrient loads from land can be assumed to be about the same here as for other countries around the Western Baltic, with the exception of the former GDR because of similar developments in sewage treatment, industry, and agriculture. Atmospheric nitrogen deposition as a more global parameter can be extrapolated from Swedish measurements (SEPA, 1990) considering an increasing tendency from central Sweden toward central Europe. For the Belt Sea the data suggest a mean annual wet deposition of about 80 mmol $m^{-2}yr^{-1}$ inorganic nitrogen (1 t km^{-2} yr^{-1}) during the 1980s.

Total runoff, nitrogen and phosphate load, and the N/P ratio of loads from Schleswig-Holstein are displayed

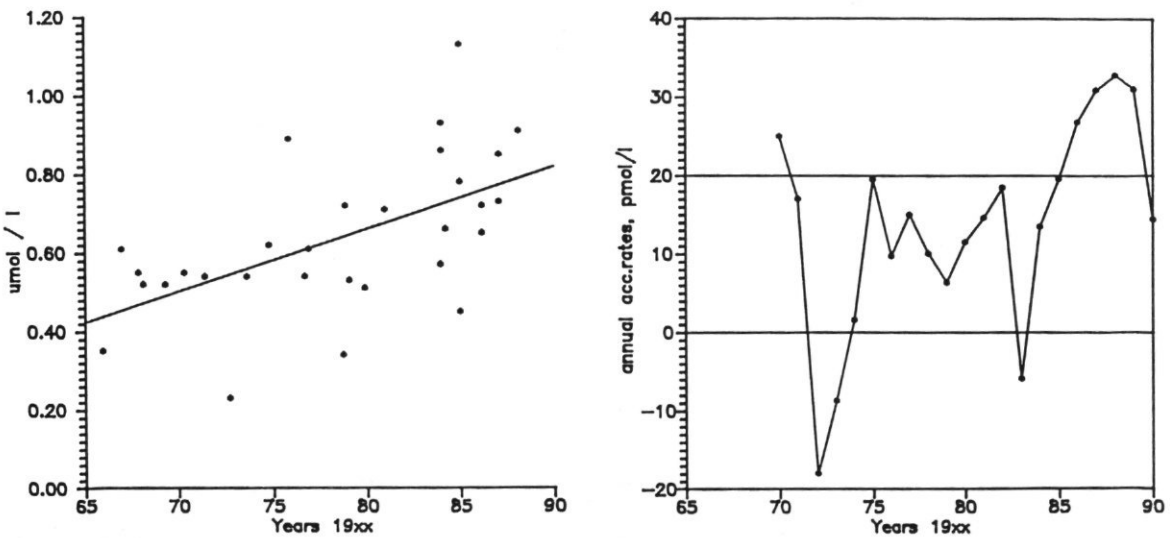

Figure 4. (Left) Phosphate daily means (Station Fladen). (Right) Trends of preceding 10 years.

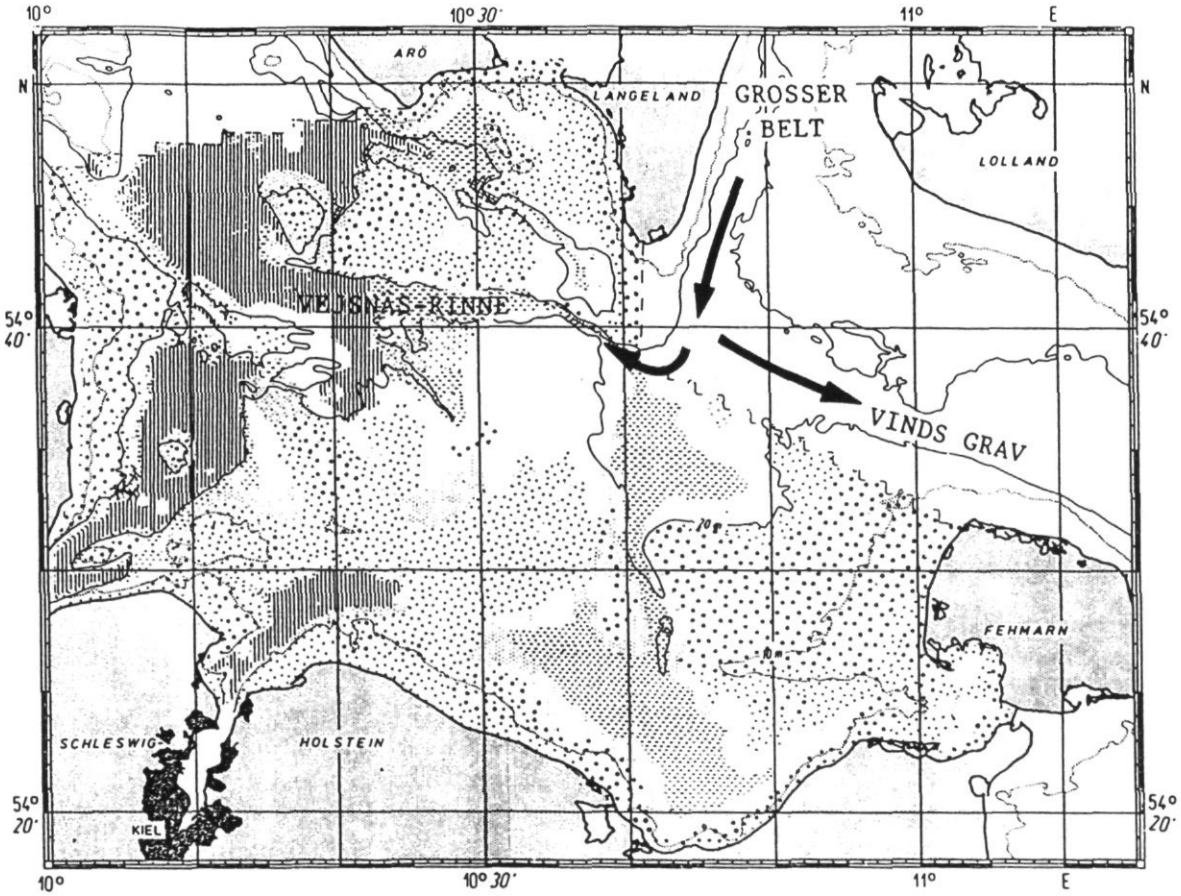

Figure 5. The Kiel Bight. Sediments: mud = shaded, sand = points. The decreasing density of points corresponds to decreasing mud contents in the sand (after Balzer, 1985).

in Figure 6a–c (LWK, 1990). About 56% of this discharge enters the Kiel Bight. There is a decrease in nitrogen load (Fig. 6a) which is, however, caused mainly by one extremely high value in 1981 and an extremely low value in 1989, both coinciding with corresponding signals in runoff (rain).

It is very difficult to calculate the atmospheric contribution to the nitrate load in the 1980s. The measurements show high year-to-year variations and a statistically significant trend cannot be calculated. The phosphate load decreases significantly from about 1000 in 1980 to 300 t yr^{-1} in 1989, obviously reflecting the reduction of phosphate contents in washing powders.

A trend in primary production during spring bloom cannot be established. The peak signal is too short to be represented by the few samples. The primary production values during the summers of 1986 to 1988 were nearly twice as high as during 1979 to 1984, while chlorophyll *a* did not increase. This may be interpreted as increasing "new production" relatively to "regenerated production" due to eutrophication or changes in vertical exchange processes. The small trends in salinity (Table 2) resulting in reduced salinity differences between surface and bottom waters could stimulate increased vertical exchange; however, the trends are only weak. Other possible explanations are increased grazing or variations in species composition modifying the assimilation ratio (BIOMON, 1990).

There are some additional indications of biological changes during the second half of the 1980s. Some species (*Porocentrum minimum* and *Heterocapsa triquetra*) which have not been reported before 1985 are now found regularly in sometimes high concentrations in some of the fjords. Since 1983 blooms of *Dictyocha (Distephanus) speculum* are reported during April and May. A remarkable phenomenon is that this species found in the Kiel Bight does not develop the siliceous skeletons which normally characterize the genus *Dictyocha*. This may be intrepreted as a result of environmental changes (Jochem and Babenerd, 1989).

The sub-halocline zoobenthos community has clearly suffered from the oxygen deficit in 1981. The species composition has changed in a way that long-living species vanish for several years while short-living, highly productive species of usually low biomass start to dominate (Weigelt, 1987; Weigelt and Rumohr, 1986). Low

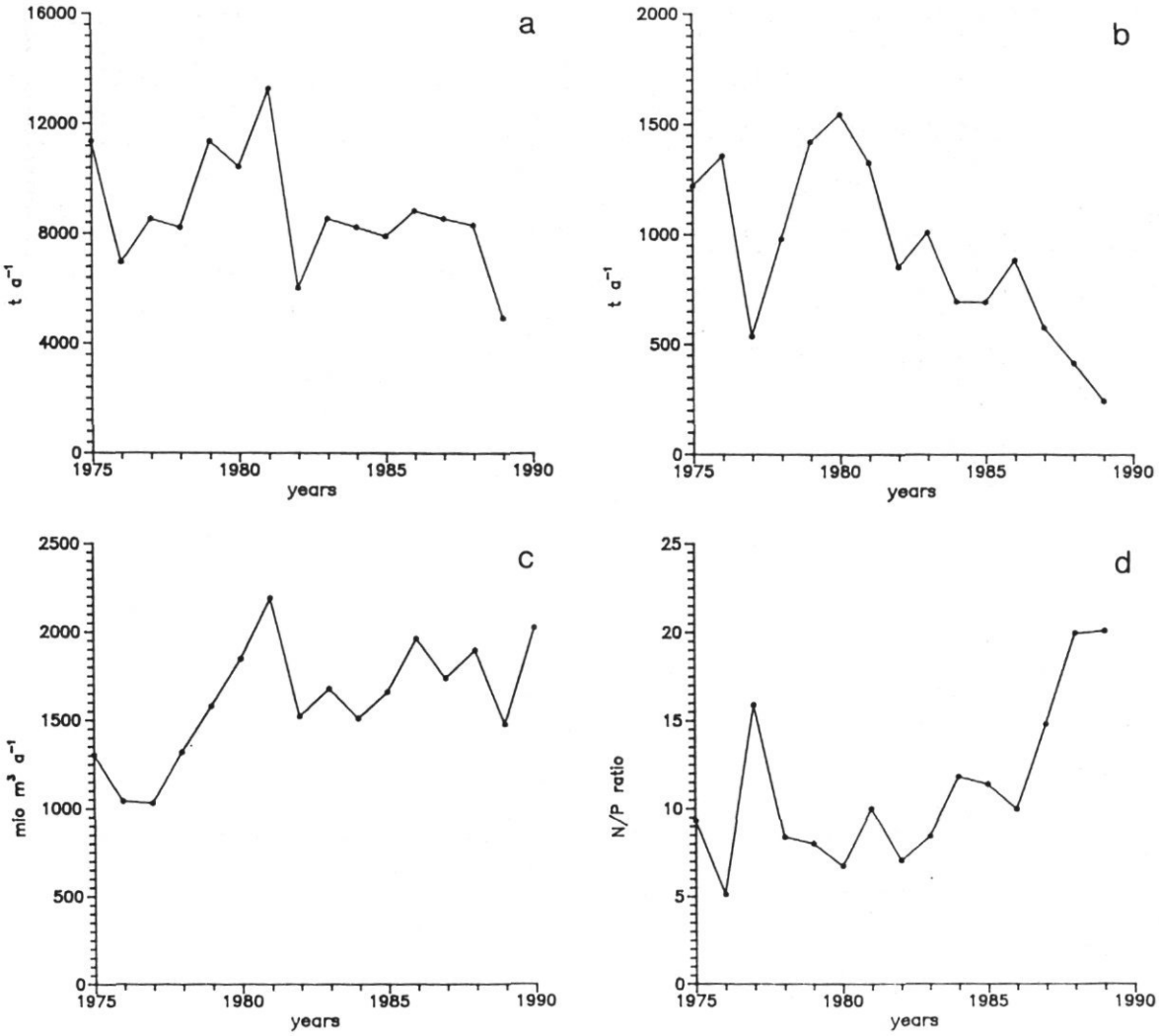

Figure 6. Loads and runoff to the Baltic Sea from Schleswig-Holstein, 1975–1989. (a) Inorg. nitrogen, (b) phosphate, (c) total runoff, (d) N/P ratio.

oxygen conditions in the Kiel Bight until 1989 have retarded the recovery of the benthic community.

Conclusions and discussion

The computation of the nutrient loads as well as the nutrient concentration measurements in the winter waters suggest that conditions in the Kiel Bight have not changed toward higher eutrophication within the 1980s. The symptoms of eutrophication, however, e.g. primary production, oxygen deficits, negative development of benthic fauna, are contradictory.

The Baltic Sea nutrient budget model of Wulff and Stigebrandt (1989) allows us to check the relation between nutrient load and observed concentrations in winter surface waters. These authors assume increasing loads for P until 1975 and for N until 1980 and constant loads since then. They predict a continuing weak increase in nitrogen concentrations until 1980 and a nearly constant level after this time, while the phosphorus concentrations will increase significantly until the end of the century.

The model fits quite well with respect to nitrogen. The nitrogen load from land is slightly decreasing but compensated by still increasing atmospheric loads. The concentrations of total N in winter surface waters are increasing by about 1 μmol $dm^{-3}yr^{-1}$.

For phosphorus loads it seems that the Western Baltic area is slightly better off than the Baltic Proper. The loads from land have been decreasing significantly (to about 30%). Thus the phosphorus concentrations in surface waters are no longer increasing (actually they are slightly decreasing), and, as opposed to the model prediction, the N/P ratio is not decreasing significantly.

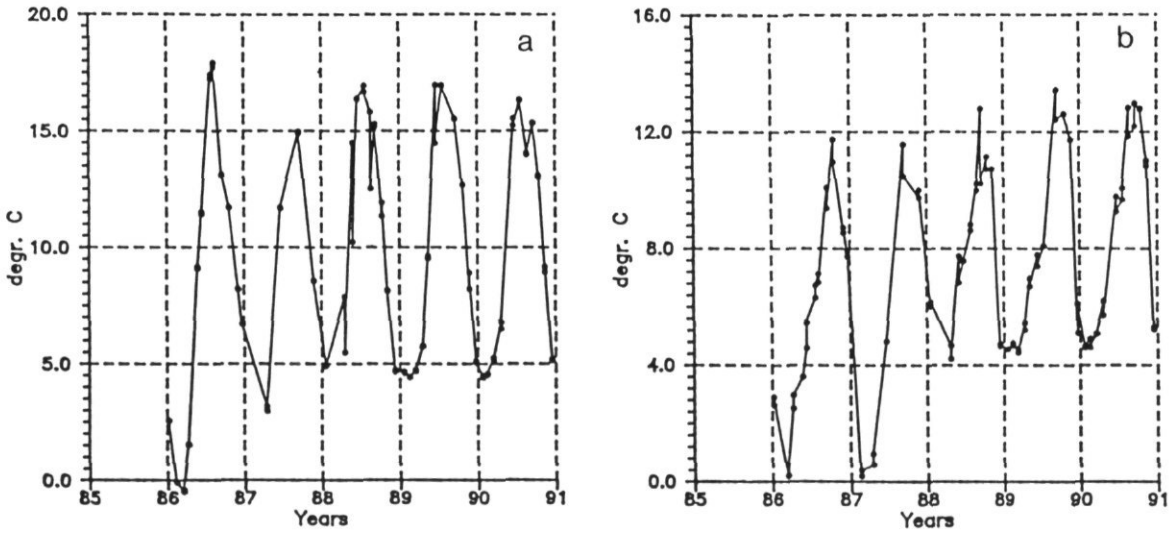

Figure 7. Temperature, station Boknis Eck, 1986–1990. (a) Surface, (b) bottom.

So the nutrient pool does not explain the "eutrophication symptoms" mentioned above and we have to look in more detail for changes within the 1980s which could possibly

- mask, produce, or simulate trends in the winter data
- affect the oxygen concentrations
- increase primary productivity during summer.

There is at least one additional factor influencing the oxygen situation: The general explanation for decreasing oxygen concentrations in bottom waters is an increased amount of organic material and the consequent oxygen demand by microbial degradation. During the second half of the 1980s, however, we have an unusual meteorological situation, which may affect the oxygen conditions even more. The last cold winter was 1985/1986. Since then the minimum water temperatures (Figs. 7a–b) have increased from about zero to 5°C. One consequence is that the oxygen solubility is 1.2 $cm^{-3}dm^{-3}$ less than in "normal" years. The additional oxygen uptake during the cooling period (mainly February) is illustrated in Figure 8b. The interannual variability corresponds with the variability of the summer minimum values (Fig. 8a).

Let us assume a small local basin (bowl) of 1 m depth which turns to stagnation during summer, which means

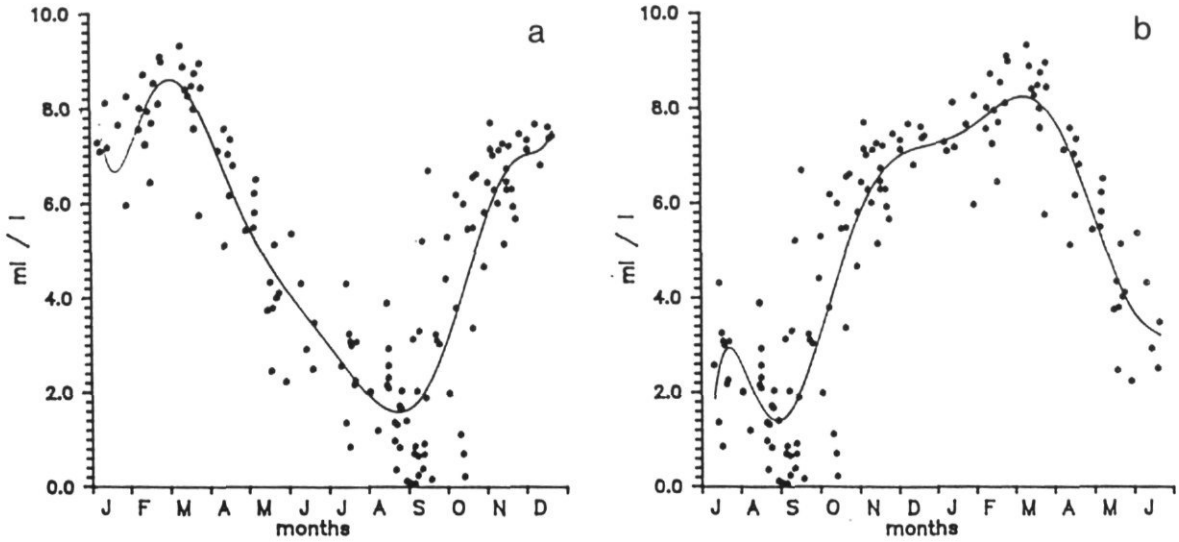

Figure 8. Means of bottom oxygen concentrations, Fehmarn Belt, 1979 to 1991. (a) January–December, (b) July–June.

that stratification restricts vertical resupply. Oxygen consumptions of sediment surfaces in the Kiel Bight are about 500 $cm^{-3}m^{-2}d^{-1}$ (Balzer, 1978; Hansen, 1980). The bottom water would become anoxic by oxygen consumption of the sediment surface after 14 days of stagnation instead of 17 days in "normal" years. Taking into consideration the additional bacterial oxygen consumption in the water column, we may find local anoxic conditions after just 8 days of stagnation (instead of 10 days). This may well explain some of the oxygen deficiency problems during the last five years.

Some additional interesting features are found in a detailed investigation of the annual fluctuations of parameters. To evaluate annual cycles we calculated polynomial optimization functions of the discrete sampling data versus the time of year. We arranged the sequence of months with respect to the time of interest, because polynomial approximations are not valid close to the data boundaries.

Figure 9a–b displays the mean annual cycles for phosphate and nitrate in surface waters of the Boknis Eck station (1986 to 1990) produced by polynomial approximation. Other methods (see e.g. Matthäus, 1971) give identical results.

The maxima – especially in phosphate – are sharper for a one-year annual cycle. One conclusion is obvious. The definition of "winter water nutrients" as the mean of January and February measurements is a very problematic compromise. As a general feature, the phosphate values (in Kiel Bight) are already decreasing after January while nitrate is still increasing until March. This phase shift is a consequence of the different remineralization rates of organic phosphorus and nitrogen. Surprising, however, is the decrease of phosphate starting as early as January. Neither primary productivity nor chlorophyll *a* confirm plankton growth at that early time of the year (main spring phytoplankton bloom is during March). This may indicate something like "luxury uptake" of nutrients. However, I will leave this discussion to the biologists. A second finding which I cannot explain at this moment is a systematic delay of the occurrence of the maxima in both phosphate and nitrate since 1979 (Fig. 10a–b). This may be a consequence of either biological or meteorological changes, but it could also be caused by changes in the nutrient load (time of release) from agriculture.

Here I would like to present a curiosity regarding the sampling in the Kiel Bight. For good scientific reasons all laboratories try to service "their" stations at regular (mostly monthly) intervals. For less scientific reasons nobody wants to sample just after New Year's day. The result is displayed in Figure 11.

Sampling at monitoring stations is concentrated around the middle of the month. We should remember that statistical treatment of data aiming at the identification of systematic concentration trends assumes that the data either represent the maximum peak or are distributed randomly over the maximum range. If not, statistical calculations may produce artificial results which could indicate accumulation or decrease but actually only reflect non-random properties of the sampling.

All these findings strongly support the idea of a change of sampling strategies for trend and budget calculations as suggested for example by Wulff and Stigebrandt. It seems much more promising to concentrate on a few representative stations which are serviced sufficiently often to access the annual cycles and identify, quantify, and eliminate short-term extrema caused

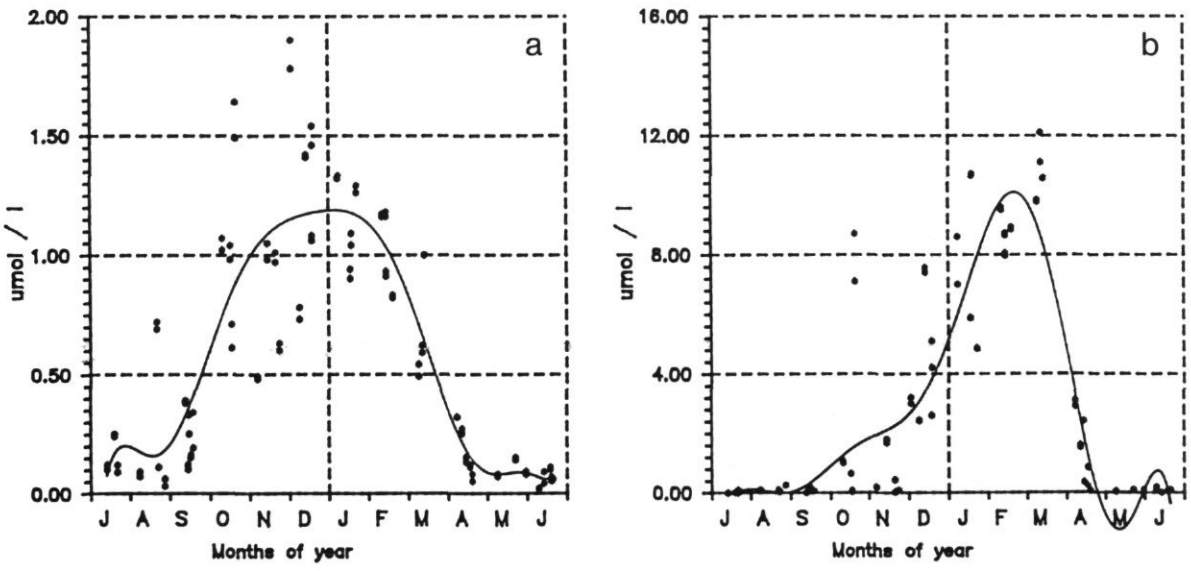

Figure 9. Mean annual cycles, surface, Boknis Eck, 1986–1990. (a) Phosphate, (b) nitrate.

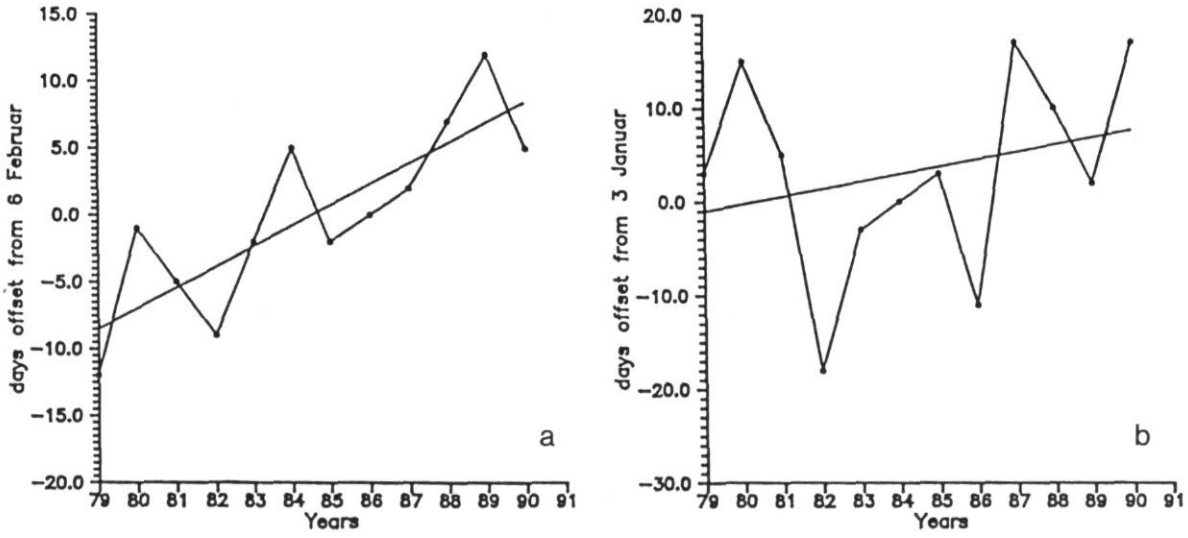

Figure 10. Shift of surface winter maxima (Boknis Eck). (a) Phosphate, days rel. to 4 January, (b) nitrate, days rel. to 8 February.

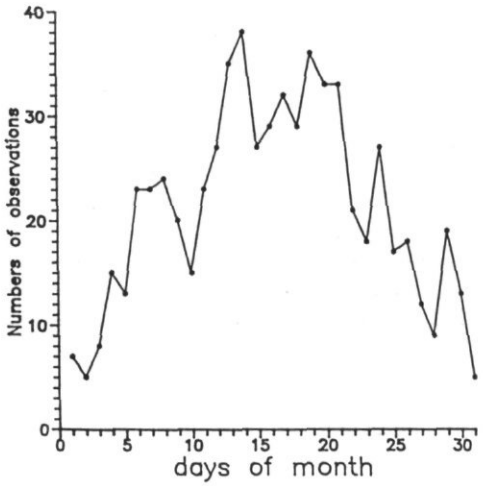

Figure 11. Station service vs. days of month (all months), Kiel Bight, 1979–1990.

by unusual events (loads, meteorological events, etc.). Long-term trends may be more reliably calculated from changes in the real winter maxima (or summer minima in the case of oxygen) as evaluated from the annual cycles instead of means of a fixed time range of sampling.

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Long-term observations of oxygen in the Skagerrak

Einar Dahl and Didrik S. Danielssen

Dahl, E., and Danielssen, D. S. 1992. Long-term observations of oxygen in the Skagerrak. – ICES mar. Sci. Symp., 195: 455–461.

The Skagerrak has a deep basin of about 700 m and a threshold of about 270 m to the Norwegian Sea. From the transect Torungen (Norway)–Hirtshals (Denmark), Flødevigen Marine Research Station has been collecting hydrographical data since 1951. Although no continuous series of oxygen measurements has been undertaken since 1951, scattered data on oxygen were obtained from the deep part of the Skagerrak in the 1950s, no data in the 1960s, and continuous data from 300-m depth and below since 1972. Since 1983, oxygen measurements from all standard depths at the more shallow stations outside Hirtshals have also been included in the programme. The oxygen conditions at the 500-m and 600-m depths in the Skagerrak were stable, with fluctuations of oxygen concentrations within the range 5.5–7.3 ml l^{-1} throughout the period 1951–1986. After reaching a peak of more than 7 ml l^{-1} of oxygen in April 1986 the concentrations have shown a downward trend, reaching less than 5 ml l^{-1} in 1990, the lowest oxygen concentration recorded in the Skagerrak deep water for the whole period 1951–1990. This was caused by a relatively long period of poor deep-water exchange during the preceding years owing to unusually high winter temperatures. The period of poor water exchange was characterized by decreasing density and increasing temperature and nutrient values in the deep water. During the winter and spring of 1991 a renewal of the deep water recurred. The oxygen measurements near the bottom in the shallow areas near the Danish coast have not, since 1983, revealed such a serious deficiency as has occasionally been observed in the North Sea west of the Danish coast and in the Kattegat in the 1980s. The fluctuations in oxygen concentrations have been in the range 4.5–7.5 ml l^{-1} and have shown a seasonal variation, with the lowest concentrations in late summer/early autumn.

E. Dahl and D. S. Danielssen: Institute of Marine Research, Flødevigen Marine Research Station, N-4817 His, Norway.

Introduction

The water masses in the Skagerrak are influenced by water from both the North Sea and the Baltic/Kattegat. Thus the water quality of the Skagerrak is partly dependent on the water quality in these surrounding areas. In the early 1980s a few incidences of low oxygen concentration were recorded in the southeastern North Sea (see Westernhagen *et al.*, 1986). Throughout the 1980s more episodes of low oxygen concentrations in the Kattegat have had adverse effects on benthic organisms (Rosenberg and Loo, 1988) and demersal fish (Phil, 1989).

In the deep water of the Kattegat (Andersson and Rydberg, 1988; Bagge *et al.*, 1990; Olsen, 1990) and in Swedish coastal bottom waters (Rosenberg, 1990), a negative oxygen trend has been recorded in recent decades, as it has for the subsurface water below 20-m depth in the Oslofjord (Baalsrud and Magnusson, 1989).

Recent studies (Weering *et al.*, 1987) have revealed

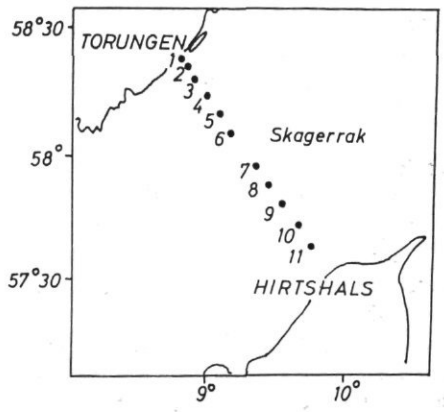


Figure 1. The transect Torungen–Hirtshals in the Skagerrak with stations 1–11.

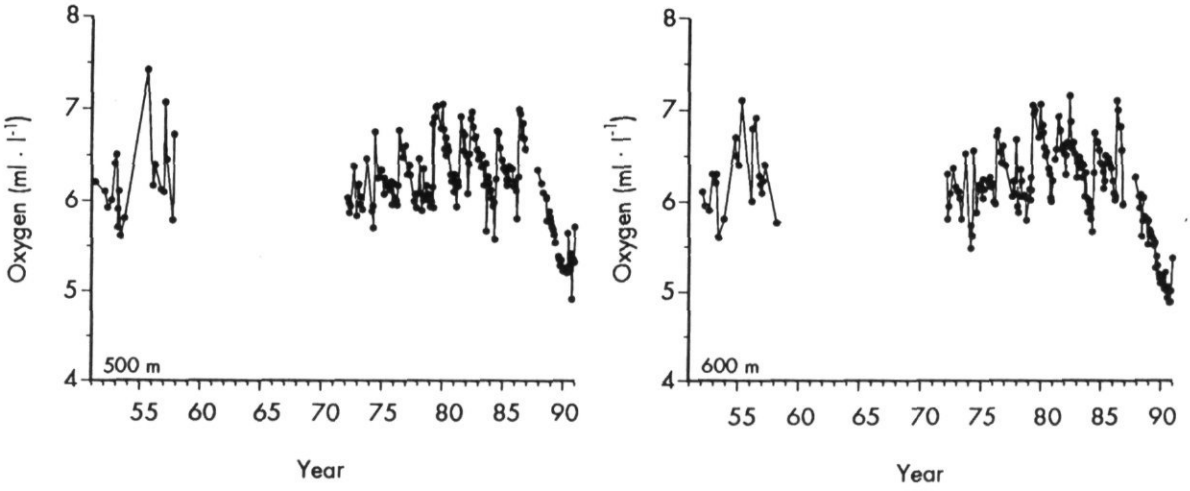


Figure 2. Oxygen concentrations (ml l^{-1}) at station 6 at 500 m (left) and 600 m (right) depth in the period 1951–1990.

that the deep part of the eastern Skagerrak is an important sediment sink, and in the same area an increase in the total benthic biomass throughout this century has been observed (Rosenberg *et al.*, 1987). Against this background, concentrations of oxygen in the Skagerrak are presented and discussed.

Materials and methods

The Skagerrak (Fig. 1) can be looked upon as part of the North Sea and as more or less a fjord with a threshold to the west of about 270 m and a central deep basin of about 700 m (Svansson, 1975). Along the transect Torungen (Norway)–Hirtshals (Denmark) (Fig. 1) hydrographical observations, with an up to monthly frequency, have been undertaken at 11 to 12 fixed stations since 1951.

No continuous series of oxygen measurements exist since 1951. However, the Flødevigen Marine Research Station has scattered data from the deep part of the Skagerrak for the period of 1951–1958, no data for the period 1959–1971, and continuous data from 300 m and below since 1972, except for 1987 when measurements were performed in October only. Since September 1983 oxygen measurements from all standard depths at the shallower stations outside Hirtshals have also been included in the programme.

From this set of data the oxygen concentrations at 500-m and 600-m depth at station 6 (58°08′N, 09°11′E) are presented together with simultaneous data on density, temperature, and salinity. In addition, the oxygen concentrations at all standard depths at station 11 (57°42′N, 09°47′E) are shown.

The oxygen was analysed using the Winkler method (Winkler, 1888) through the whole period, while the temperature was measured by Nansen reverse thermometer until 1987 and since then by CTD (Neil Brown). The salinity was analysed by titration until 1975 and thereafter by salinometer (Guildline Instrument) and since 1987 by CTD (Neil Brown). Water samples were collected by Nansen bottles until 1987 and by Niskin bottles mounted in a rosette sampler since then. From 1987 onwards one could be sure of having collected water from the correct depth with the use of the depth indicator on the CTD. Before that time one had to trust the recordings of the wire length by the meter wheel, recording the wire length out in the sea. Our experience is, however, that owing to weak currents in the central Skagerrak, the wire with the water bottles always seemed to hang vertically in the sea in this area, and thus the meter wheel gave the correct sampling depth.

From the deep part of the Skagerrak a few data on nutrients, orthophosphate, nitrate, and silicate, are given. The nutrients were analysed according to the autoanalyser system described by Føyn *et al.* (1981) with some minor changes.

Results

The oxygen conditions at the 500-m and 600-m depths of the Skagerrak reflect several periods of about 1–2 years with an oxygen decrease, indicating periods of more or less stagnation (Fig. 2). Periods with increasing oxygen concentrations, indicating renewal of the deep water, were of shorter duration, some a few months only. There was no decreasing trend throughout the period 1951–1986 (Fig. 2). The fluctuations in the latter period, both depths included, were within the range 5.5–7.3 ml l^{-1}. After a peak of more than 7 ml l^{-1} in April 1986, however, our data demonstrate a downward trend for four years, reaching less than 5 ml l^{-1} in 1990, the lowest oxygen concentrations recorded in the deep part of the Skagerrak for the whole period 1951–1990. This downward trend has probably been broken by a renewal

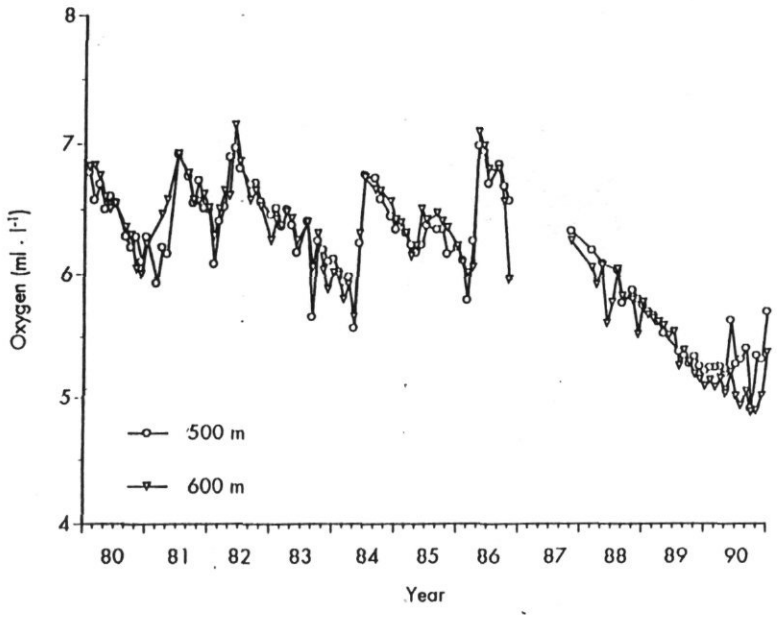


Figure 3. Oxygen concentrations (ml l^{-1}) at station 6 at 500 and 600 m depth in the period 1980–1990.

period in 1987, but unfortunately we have only few data from that year.

The more frequent and regular measurements from the 1980s show a more detailed picture of the periods of a few months' duration with rapidly increasing concentrations and periods of up to years with more slowly decreasing oxygen concentrations (Fig. 3, Table 1). The peak oxygen concentrations after renewal of the deep water, i.e. 6.5–7.2 ml l^{-1}, have always been recorded in the period April–June (Table 1).

Also concerning σ_t, temperature and salinity, the variation at 500-m and 600-m depths showed similar patterns for the period 1951–1990 (Fig. 4). Since the summer of 1986, σ_t, temperature, and salinity have been more or less continuously rising. The periods of obvious increase in oxygen at these depths, the renewal periods, are characterized by increasing σ_t owing to decreasing temperatures (Table 1).

The concentrations of the nutrients orthophosphate and nitrate in the deep part of the Skagerrak have mainly been below 1.0 μmol l^{-1} and 12 μmol l^{-1} respectively and quite stable throughout the 1980s, except for an increase to more than 1.5 μmol l^{-1} and 14 μmol l^{-1} respectively at the end of the decade (Fig. 5). The concentrations of silicate, although only analysed since 1988, show an even clearer increase (Fig. 6), and the simultaneous observations of the silicate concentrations at each depth seem to be pretty well separated.

At station 11, about 10 nautical miles off Hirtshals, the oxygen conditions for the period September 1983–December 1990 were good (Fig. 7). The lowest oxygen concentration recorded was 4.5 ml l^{-1} at 60 m depth near the bottom. The concentrations showed typical seasonal changes with the highest values in the winter and early spring and the lowest in late summer and early autumn.

Table 1. Recordings of oxygen (ml · l^{-1}) at station 6 at 600 m depth after obvious renewal and stagnation periods and the corresponding temperatures, including the approximate length of the stagnation and renewal periods and the respective decrease and increase of oxygen (ml · l^{-1}) during the periods.

End of renewal period			Stagnation period		End of stagnation period			Renewal period	
Date	O_2 (ml · l^{-1})	°C	Length	O_2 decrease	Date	O_2 (ml · l^{-1})	°C	Length	O_2 increase
June 1981	6.92	5.30	7 months	0.63	Jan 82	6.29	5.67	5 months	0.85
May 1982	7.14	4.78	23 months	1.48	Apr 84	5.66	5.87	2 months	1.08
June 1984	6.74	5.36	9 months	0.60	Mar 85	6.14	5.56	2 months	0.36
May 1985	6.50	5.08	9 months	0.49	Feb 86	6.01	5.54	2 months	1.08
Apr 1986	7.09	4.82	53 months*	2.21	Sep 90	4.88	6.04	**	

*The stagnation was probably broken during 1987, the year with recordings in October only.
**1 May 1991 the oxygen concentration at 600 m was 6.64 ml · l^{-1} and the temperature 6.40°C after a maximum recording of 7.06°C in December 1990.

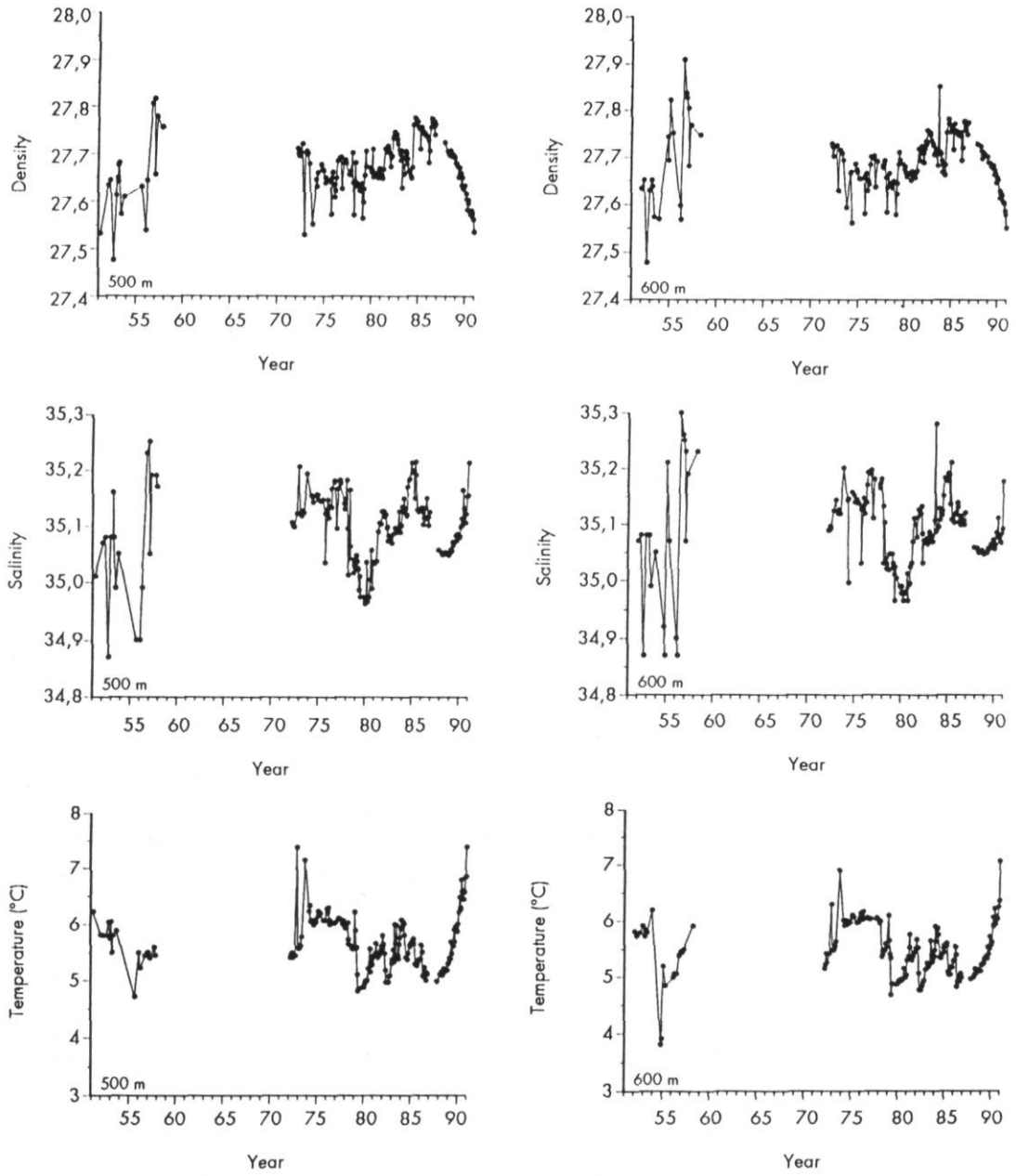
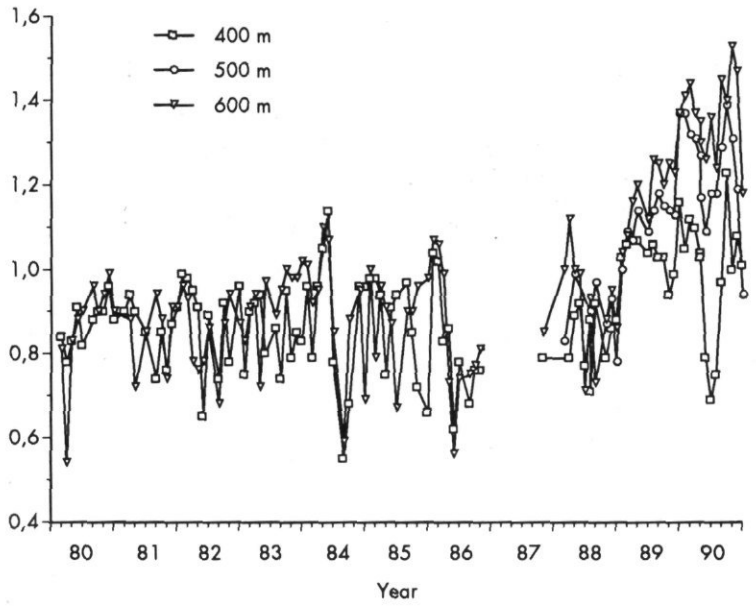


Figure 4. Density, (σ_t) salinity (PSU), and temperature (°C) at station 6 at 500 m (left) and 600 m (right) depth in the period 1951–1990.

Discussion

The changes in oxygen concentrations in the deep parts of the Skagerrak during 1951–1990 do not show any overall decreasing trend, as has been shown for some bottom waters along the Swedish west coast (Rosenberg, 1990) and in the Oslofjord (Baalsrud and Magnusson, 1989). The downward trend at the end of the 1980s is caused by a relatively long period of poor deep-water exchange. This can be judged from the more or less continuous increase of σ_t, temperature, and salinity in the same period (Fig. 3). The increase in concentrations of the nutrients, especially silicate, further support this hypothesis of poor water exchange. Figure 6 indicates that different concentrations of silicate seem to be powerful parameters for separating different degrees of stagnation in the area.

Recent data demonstrate that a deep-water exchange



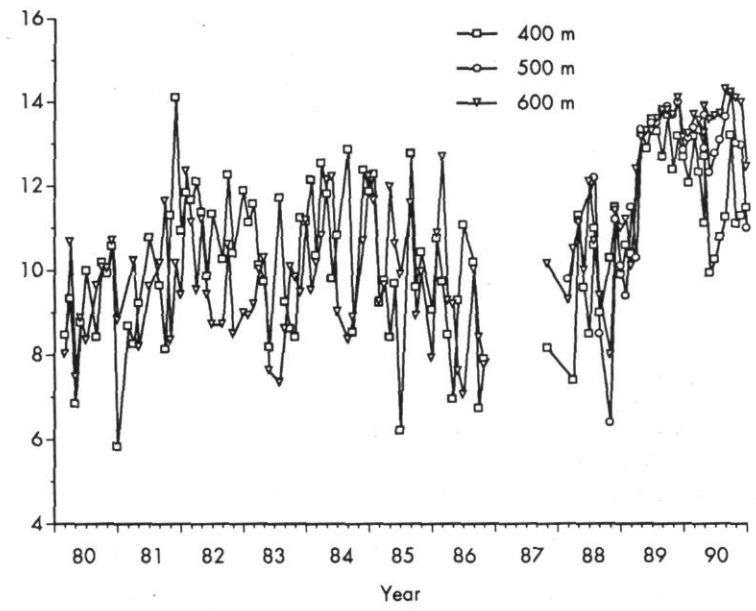


Figure 5. Orthophosphate and nitrate (μmol l^{-1}) at station 6 at 400, 500 (only few recordings), and 600 m depth in the period 1980–1990.

took place during the months of February–April 1991. The recorded temperature, salinity, and density at 600-m depth in May 1991 were 6.40°C, 35.211 PSU, and 27.678, respectively, and concentrations of nitrate, phosphate, and silicate were 11.23 μmol l^{-1}, 0.85 μmol l^{-1}, and 6.65 μmol l^{-1} (Danielssen, 1991). This further supports the idea that the downward trend of oxygen since 1986 is a periodic one due to temporary poor deep-water exchange, and demonstrates the importance of long-term and regular observations for getting reliable data for trend analysis.

According to Ljøen (1981) the more radical exchange of deep water in the Skagerrak basin is due to a cascading overflow of North Sea Shelf waters after cold winters. This water exchange results in decreased temperature and salinity at the bottom. Our data (Table 1)

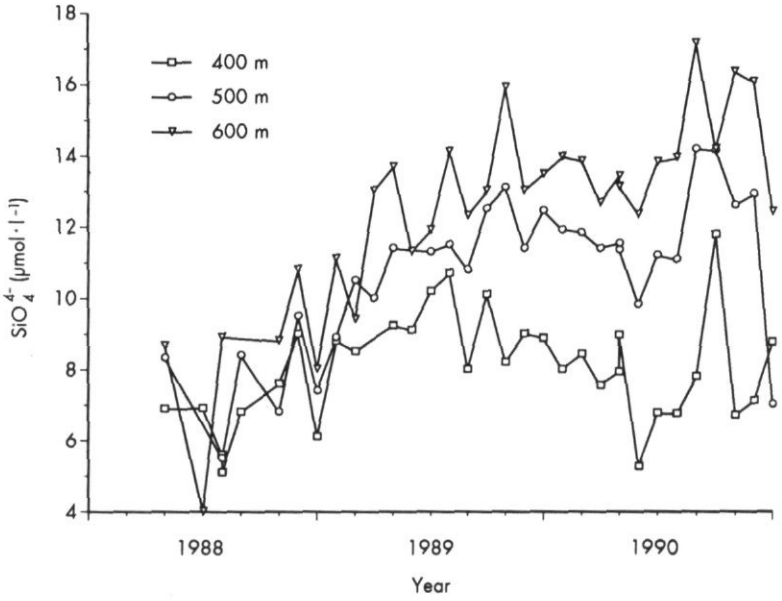


Figure 6. Silicate (μmol l^{-1}) at station 6 at 400, 500, and 600 m depth in the period 1988–1990.

demonstrate this. The mild winters in the North Sea area after 1987 have prevented such deep-water exchange in the Skagerrak and thus caused the relatively long period of stagnation throughout the last part of the 1980s.

The oxygen conditions recorded in the southern Skagerrak (station 11) since 1984 must be characterized as normal and good, and the levels and seasonal fluctuations are well in correspondence with the data for "Skagerrakwater", water of salinity more than 30 PSU, in Andersson and Rydberg (1988). No remarkable low values, as observed in the southeastern North Sea in the early 1980s (Westernhagen *et al.*, 1986), have been recorded. However, after 1983 low values have not been recorded in the latter area either (Gerlach, 1990).

There is growing concern, and some evidence, that the renewal waters to more isolated basins bordering the Skagerrak–Kattegat area, i.e. the Kiel Bay (Weigelt, 1990) and the Oslofjord (Baalsrud and Magnusson, 1989), may have had decreasing quality oxygen content in recent years. If this is the case, our data indicate that this is due more to increased local oxygen consumption in the transition areas, the southern Kattegat, and the Belts and the outer Oslofjord, than to any large-scale oxygen decrease in the Skagerrak area.

Our time series from the Danish coast is too short for drawing definite conclusions concerning possible trends in oxygen content in the shallow waters near the Danish coast. The results indicate, however, that in this respect one should be especially aware of the oxygen levels in late summer/early autumn, the season when the minimum values occur.

Acknowledgments

Technicians Svein Erik Enersen, Vesla Fosback, Knut Hansen, Terje Jåvold, Kristian Kristiansen, Ernst O. Maløen, Øystein Paulsen, and Aadne Sollie carefully carried out the practical work in the field and in the laboratory. The crew on board the research vessel G.M. "Dannevig" are also thanked.

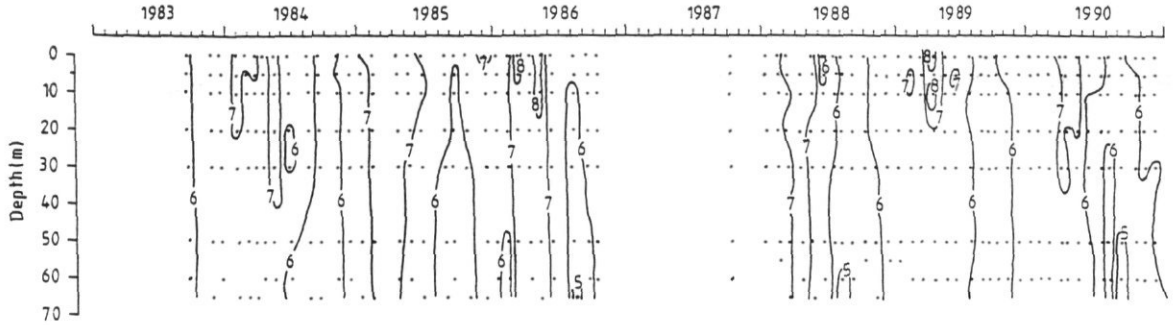


Figure 7. Oxygen (ml l^{-1}), at station 11 in the period 1983–1990.

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Variability of nutrients and chlorophyll in the Skagerrak, 1980–1990

Didrik S. Danielssen and Einar Dahl

Since the beginning of 1980 the nutrients orthophosphate, nitrate and nitrite, and chlorophyll *a* have been measured about once a month along the hydrographical section Torungen (Norway)–Hirtshals (Denmark) in the Skagerrak. Measurements of silicate have also been made, but for a shorter period. The variability is demonstrated here by the presentation of data from selected stations (Fig. 1).

Both orthophosphate and nitrate show typical seasonal variations with a tendency to higher winter values near the Danish coast than at the Norwegian coast, especially in the first half of the 1980s. There seems not to have been any tendency to change in the general picture during the whole period (Figs 2 and 3).

In the middle of the Skagerrak the surface layer also shows the same typical seasonal variations with no changes in the general picture during the whole period. In contrast to observations at the Danish and Norwegian coasts, high orthophosphate and nitrate values are recorded every year at 20 m depth and also at other times of the year than winter, which is due to the "dome" effect.

In contrast to orthophosphate, very high values of nitrate are sometimes found near the Danish coast in the

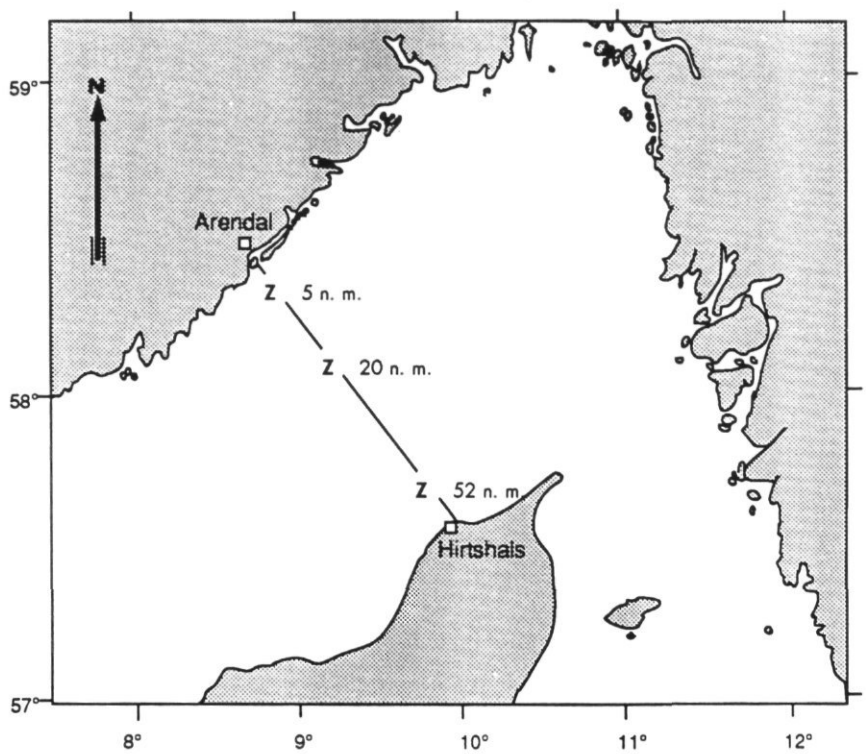

Figure 1. Map of the Skagerrak showing the positions of the stations mentioned.

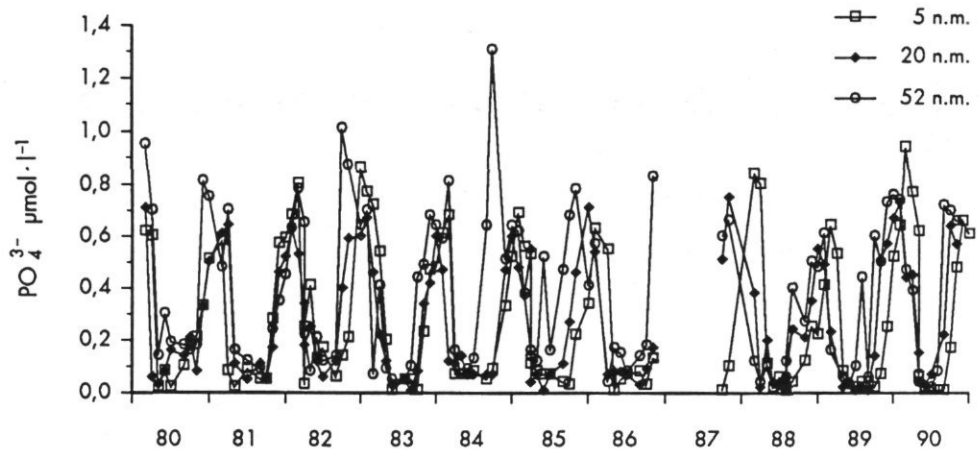

Figure 2. Seasonal variations of orthophosphate at 10-m depth at three stations along the hydrographical section Torungen–Hirtshals during the period 1980–1990.

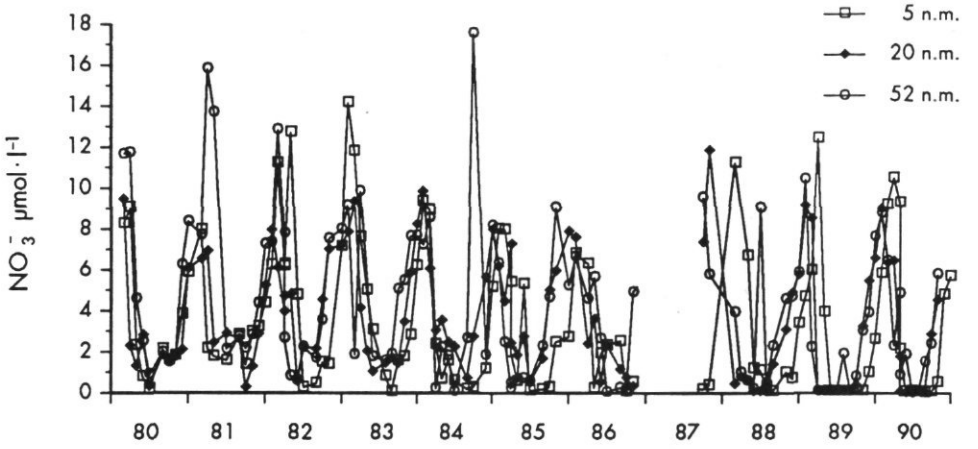

Figure 3. Seasonal variations of nitrate at 10-m depth at three stations along the hydrographical section Torungen–Hirtshals during the period 1980–1990.

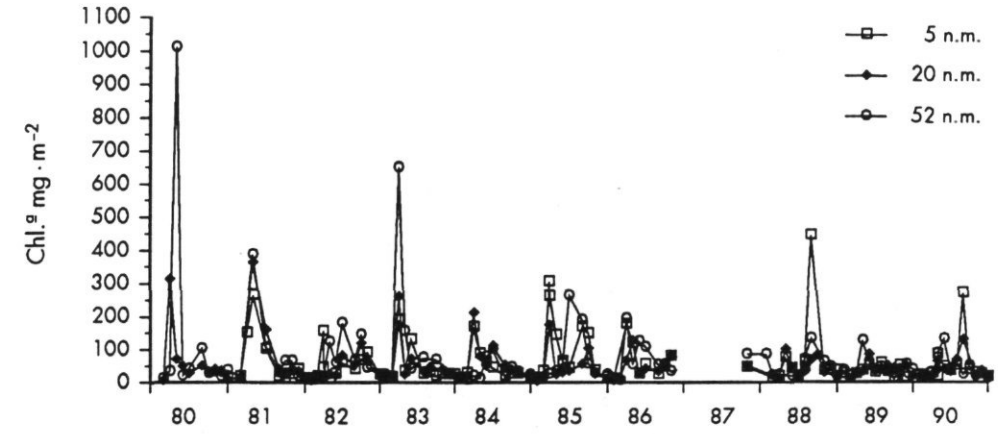

Figure 4. Seasonal variations of integrated values between 0- and 30-m depth of chlorophyll *a* at three stations along the hydrographical section Torungen–Hirtshals during the period 1980–1990.

period December–April in salinities between about 31.5 and 33.5 PSU, which could indicate water masses from the Jutland current possibly sometimes mixed with other low salinity water masses from the Kattegat. Salinities less than about 31.5 PSU always had a low nitrate content, and were observed in the summer months. Water masses with salinities lower than about 33.5 PSU always had a low silicate content and were observed during summer.

The spring bloom seemed to be much more pronounced in the first half of the 1980s than during the second half with unusual temperature conditions in the winter period (Fig. 4).

Conclusions

No trends in the amount of orthophosphate and nitrate were observed during the period investigated. Very high nitrate concentrations were sometimes observed in salinities between 33.5 and 31.5 PSU near the Danish coast. The spring blooms in the last years of the decade have been much smaller than earlier, possibly due to the unusual winter temperature conditions.

Didrik S. Danielssen and Einar Dahl: Institute of Marine Research, Flødevigen Marine Research Station, N-4817 His, Norway.

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A short note on the exchange of water and nutrients between the Skagerrak and the Kattegat

Lars Andersson

Andersson, L. 1992. A short note on the exchange of water and nutrients between the Skagerrak and the Kattegat. – ICES mar. Sci. Symp., 195: 465–473.

Simultaneous measurements of velocities, salinity, oxygen, and nutrients were undertaken in the northern Kattegat, in a cross-section between the island of Læsø and Sweden. The main aim was to determine the amount of deep water flowing southwards through the Kattegat, as a part of the exchange of water between the Baltic and the ocean. Fluxes of water, nutrients, and oxygen are presented as functions of salinity. A division is made between the biologically passive winter period (November to February), when inorganic nutrients are nearly conservative, and a "summer" period from March to October. The mean flow of deep water (S > 30 PSU) was, during winter, 64 000 m^3/s and, during summer, 28 000 m^3/s, with a yearly mean value of 40 000 m^3/s. The values are approximately 30% higher than the indirectly calculated fluxes, based on entrainment theory. The yearly mean outflow of surface water (S < 30 PSU) was approximately 20 000 m^3/s, indicating a mean outflow of more than 30 000 m^3/s in the shallow section west of Læsø. Salt balance calculations and old lightvessel velocity measurements confirm this conclusion.

Lars Andersson: *SMHI Oceanographic Laboratory, Box 2212, S-403 14 Gothenburg, Sweden.*

Introduction

The Kattegat forms the outer part of the Baltic estuary (Fig. 1). It is shallow, with a mean depth of 20 m, and is characterized by a very strong vertical salinity stratification, caused by an outflow of low saline Baltic water in the surface layer and an inflow of high saline water from the Skagerrak at deeper levels (Fig. 2). The surface water salinity varies between 12 and 25 PSU, the deep water salinity between 32 and 34 PSU. The halocline depth varies both in time and space, with an average of about 15 m.

Although the local nutrient supply to the Kattegat is large, the supply of nutrients and oxygen from surrounding waters is also important for the ecology of the area. In fact, the supply of deep water from the Skagerrak, driven by upward entrainment due to local winds (Stigebrandt, 1983) is the largest single nutrient source (Rydberg *et al.*, 1990).

In 1975–1977 the Fishery Board of Sweden carried out an extensive field programme based on current and nutrient measurements in the northern Kattegat (the GF-section, see Fig. 1). The results indicated large current fluctuations in time and space. The calculated net flux exceeded the supply of freshwater to the Baltic by a factor of three (Szaron, 1979; Svansson, 1984). The reason for this discrepancy was not quite clear, but the section coincided with a front between the Kattegat surface water and the high salinity water in the Jutland current flowing eastward north of Skagen.

In this paper, the results from an intensive measuring programme which was carried out in a cross-section near Læsø (Fig. 3) in order to determine the fluxes of water, nutrients, and oxygen into the Kattegat are presented. The programme comprised simultaneous measurements on a total of 38 occasions between 1984 and 1988. To avoid the frontal area in the northern Kattegat, the cross-section in this project was chosen further south. As our main interest concerned the deep water, this section seemed more suitable, although we lost a part of the northernmost Kattegat.

Field work

Out of the 38 expeditions, 27 were considered as "full cross-sections", including all 10 stations and without any missing parameters. The results from these 27 sections comprise the database for this report. CTD profiling with a Neil Brown, MK III, was made at all stations.

Current measurements with pendulum current meters were carried out at stations 2, 4, 6, 8, and 10. Water

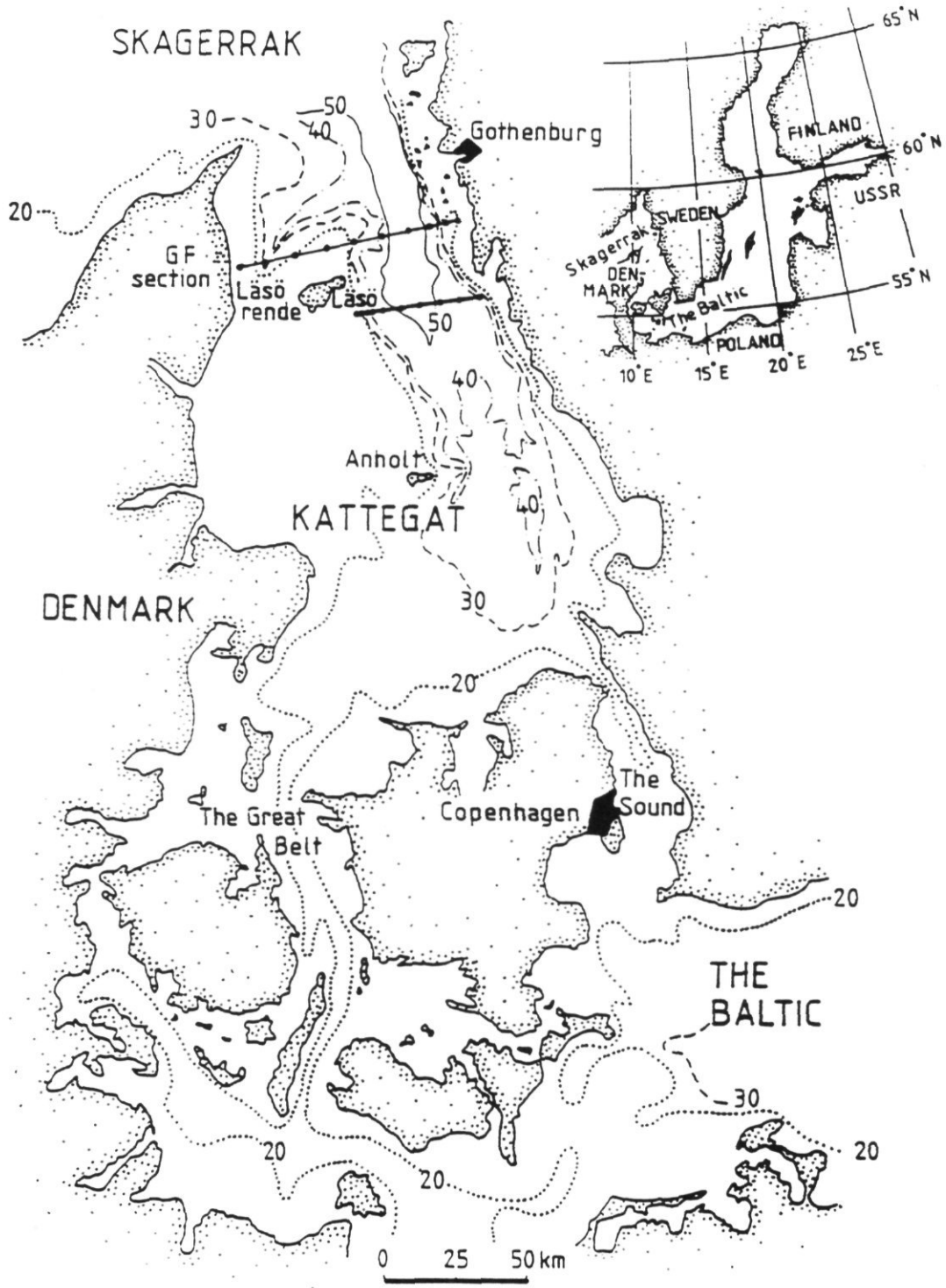

Figure 1. Map of the Kattegat area, with the Læsø and GF sections indicated.

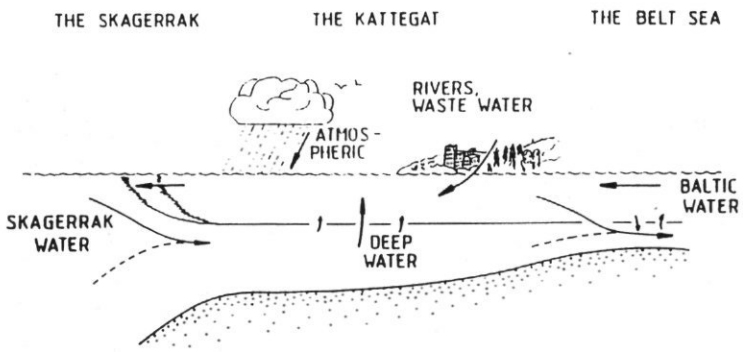

Figure 2. Sketch of a length section through the Kattegat indicating the general circulation.

sampling for nutrient chemistry, salinity, and oxygen determination was carried out at every second station (2, 4, 6, 8, and 10) using a rosette water sampler attached to the CTD. The sampling depths were determined in correspondence with the CTD profile, so that at least one sample was taken in each water mass. Between three and eight samples were taken at each station. Nutrient chemistry included determination of NO_3, NO_2, NH_4, (NO_3 + NO_2 + NH_4 = DIN), TOTN, PO_4=(DIP), and TOTP.

Data treatment

The cross-section was divided into a grid net according to Figure 4. Each grid box was assigned a mean salinity and a mean temperature by averaging all salinity and temperature data from the CTD measurements (>10 values/m) within that box.

The results from the pendulum current meters were treated as follows: The velocity recordings from the individual pendulums were divided into cross- and lengthwise components. Each box where there was a current measurement was assigned that specific value. For the boxes in between, the velocity values were interpolated using the salinity profile, assuming that the gradients in velocity were located at the same depths as the gradients in salinity.

All chemical parameters were interpolated in the same way as the velocity data, such that each box was assigned a specific concentration. This procedure resulted in one data matrix for each parameter (velocity, salinity, temperature, oxygen, DIP, TOTP, DIN, TOTN) and each transect. The mean velocity V_{ij} and mean salinity $\bar{S}_{ij}$ in each grid box were then calculated for the above-mentioned matrixes. Averages including standard deviations are shown in Figures 5a, b and 6a, b respectively.

The volume fluxes, M(S), perpendicular to the cross-section were calculated as functions of salinity (see Walin, 1977). M(S) is the volume flux for salinities less than S. M was determined in the following way: the velocity, V_{ij} in each box was multiplied by the corresponding area, a_{ij}, and sorted after salinity in 0.5 PSU intervals. Thus, we got a function.

$$m(S) = \sum_{ij} V_{ij} * a_{ij}(S)$$

where m is the volume flux in the salinity interval between S and S+ΔS and where ΔS=0.5 PSU. M was then obtained by integrating m;

$$M(S) = \int_0^S m(S')\, ds'$$

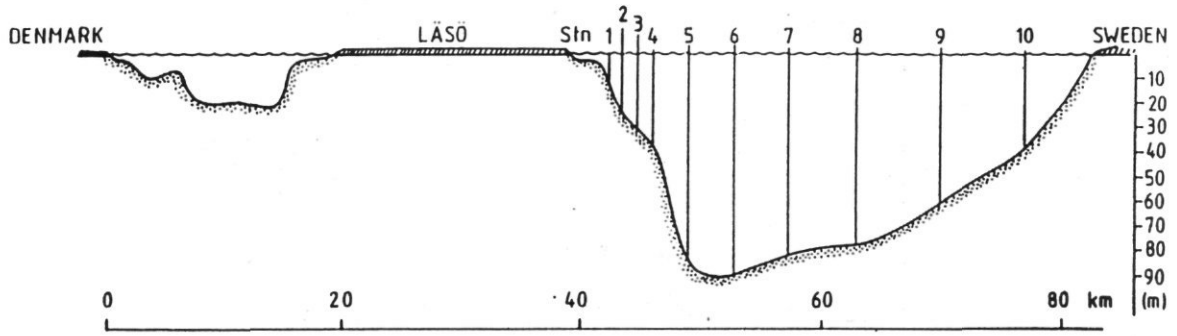

Figure 3. Cross-section from Denmark to Sweden. Locations of the stations in the Læsø section are shown.

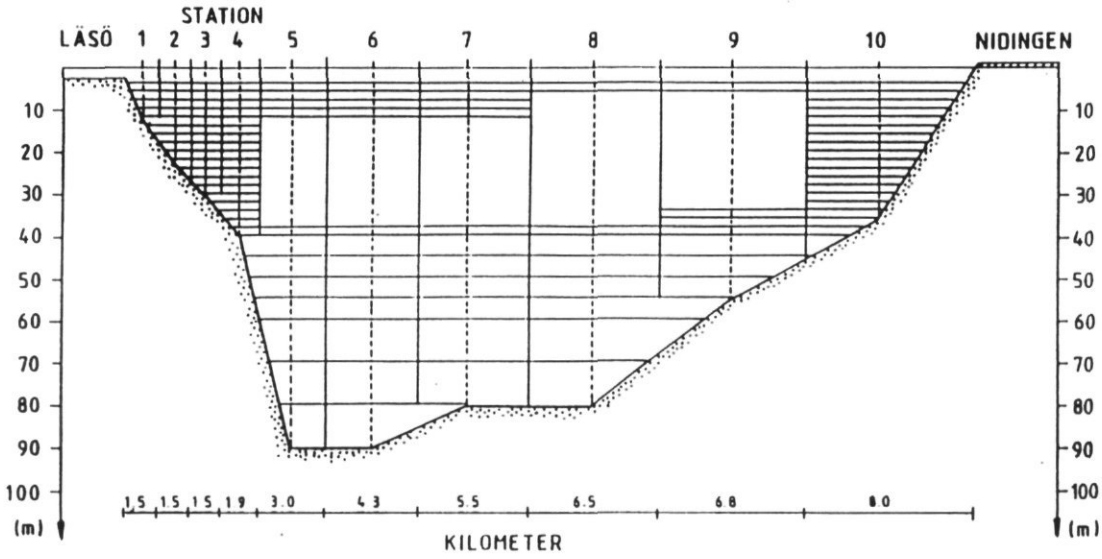

Figure 4. The Læsø section, including the grid-net used in the calculations.

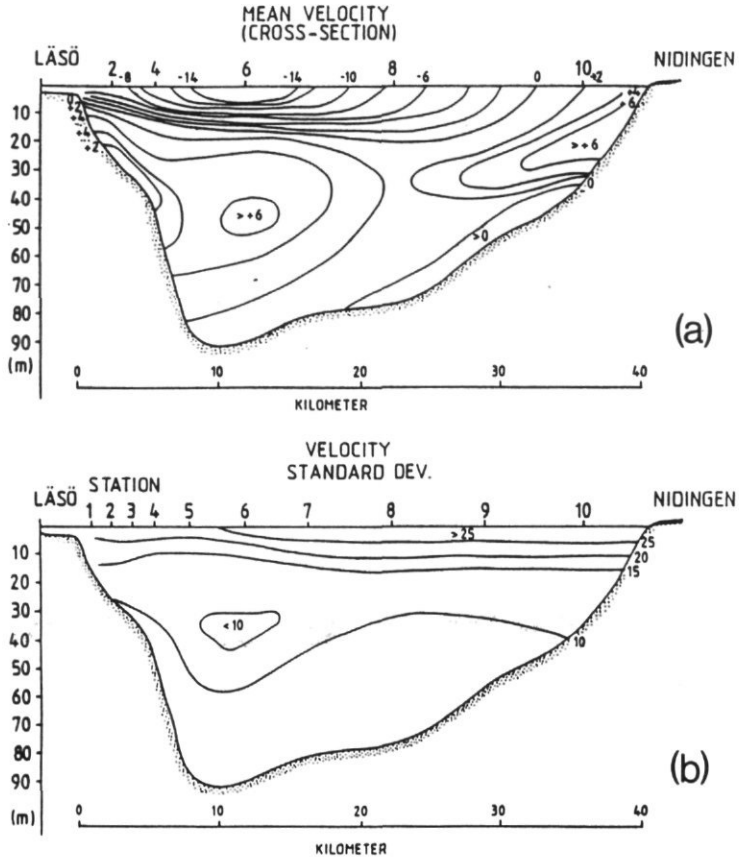

Figure 5. (a) The mean velocity (cm/s) (PSU) (mean for all transects). (b) The standard deviation of velocity.

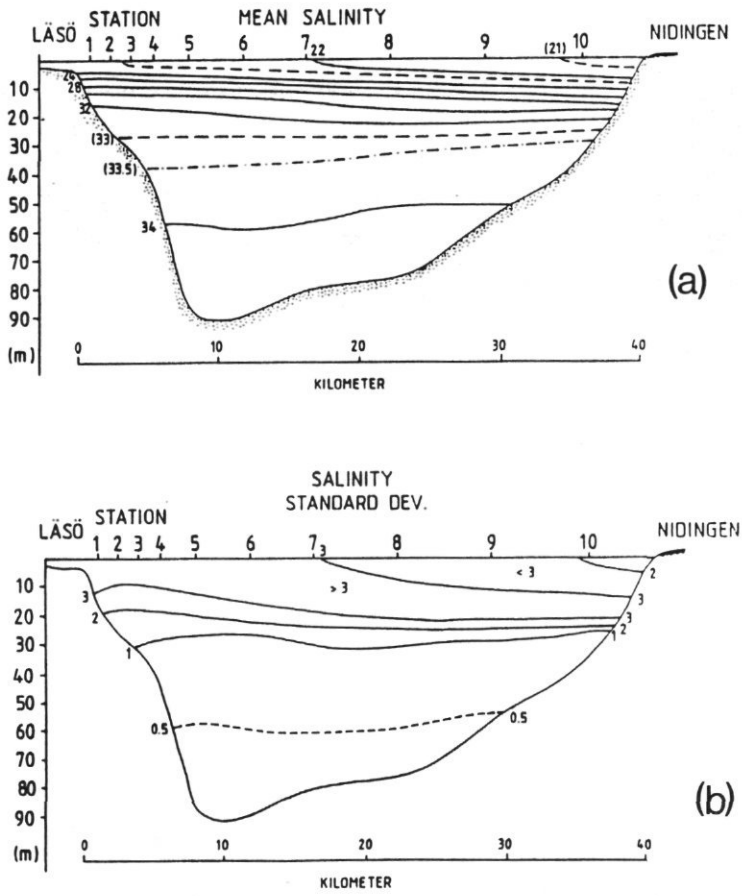

Figure 6. (a) The mean salinity (PSU) (mean for all transects). (b) The standard deviation of salinity.

These functions were calculated for each transect (see Rydberg and Andersson, 1989) and averaged over winter (Nov–Feb) and summer (Mar–Oct) periods. The average M(S) for winter and summer is shown in Figure 7a, b. The volume transport M(S), average for the whole year is given in Figure 7c.

To calculate the integrated fluxes of salt and nutrients through the section, for each crossing, we used the following expressions:

$$M_s(S) = \int_0^S v_{ij} * a_{ij} * S'_{ij}\, ds'$$

for salinity transport, and

$$M_c(S) = \int_0^S v_{ij} * a_{ij} * C_{ij}\, ds'$$

for transport of nutrients.

Results

Salinities. Fluxes of water and salt

The approximate net fluxes of water are shown in Figure 8; the mean outflow in the surface water is in the order of 18 000 m^3/s ($S < 30$ PSU), while the inflow of deep water is in the order of 40 000 m^3/s ($S > 30$). Thus, by assuming that the net outflow of water from the Baltic is 15 000 m^3/s (equal to the mean fresh water supply; see Svansson, 1975), our measurements indicate that there must be an outflow on the western side of Læsø which is well above 30 000 m^3/s.

A very strong seasonal variability was found in the volume fluxes across the Læsø section, mainly due to higher windstress and subsequent entrainment transport during the winter period.

Fluxes of nutrient and oxygen

The period from November to February coincides with

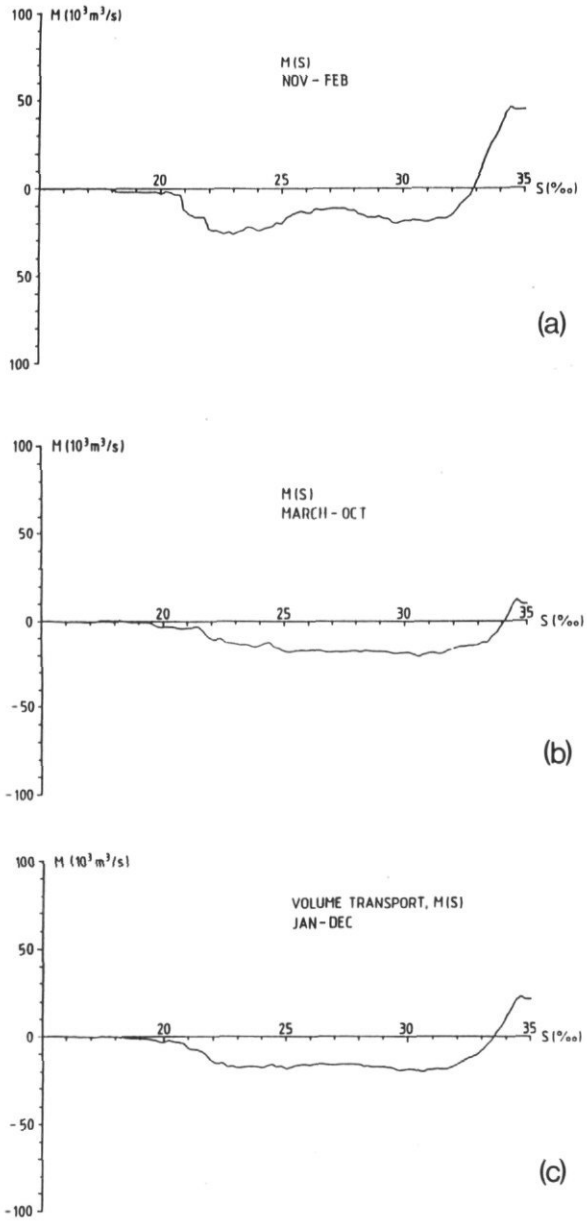

Figure 7. (a) M(S) winter. Positive values indicate transports into the Kattegat. (b) M(S) for all summer transects. Positive values indicate transports into the Kattegat. (c) M(S) whole year. Positive values indicate transports into the Kattegat.

the biologically passive season, where the inorganic nutrient components are almost conservative. From March to October, on the other hand, we can expect low concentrations of inorganic nutrients within the surface waters.

The nutrient (DIP, TOTP, DIN, TOTN) fluxes through the cross-section are shown in Figures 9a–d (winter) and 10a–d (summer). The flux of DIN into the Kattegat within the deep water during winter is 24 000 tons N/month (S > 30 PSU), but during summer it is only 6000 tons N/month (S > 30 PSU), owing to a combination of less inflow of deep water and lower concentrations. In the surface water, the outflow during winter is c. 6000 tons N/month (S < 30 PSU), while during summer the surface water outflow is negligible. The corresponding fluxes of DIP in the deep water during winter are 4000 tons P/month and during summer 1100 tons P/month. The surface water outflow during winter is 800 tons P/month and during summer about 200 tons P/month. Thus, differing from DIN, there is a small export of DIP from the Kattegat during summer.

As the concentrations of TOTN and TOTP are relatively constant throughout the year (with only a small decrease in concentrations during summer), the transport of these constituents more directly mirrors the volume transport. During winter, however, there is a net transport of TOTN into the Kattegat of 23 000 tons N/month. The corresponding inflow of TOTP is 4 000 tons P/month. Discussing net fluxes, however, one must keep in mind that the fluxes west of Læsø are not considered in these figures. Taking this transport into consideration TOTN and TOTP are transported out of the Kattegat.

The oxygen deficit and the supply of oxygen to the Kattegat deep water is by far the most important question for the ecology of the area. Figure 11 a–c shows the fluxes of oxygen in the Læsø section. During winter the deep-water inflow is large. The oxygen concentrations in the Kattegat deep water approaches those of the surface water (Andersson and Rydberg, 1988). During summer, the supply of oxygen through the deep water is still large, and even during the period when the oxygen concentrations within the Kattegat deep water are really low (August–October), the supply is still large; more than 500 000 tons O_2/month (S>30 PSU).

Discussion

This project was carried out with the main purpose of determining the flux of high saline (S > 30 PSU) Skagerrak water, including its salt, nutrient, and oxygen content, into the Kattegat deep water. However, measurements were taken also within the surface water, excluding only the rather small and shallow section west of Læsø (Læsø Rende, Fig. 3). It would otherwise have been impossible for us to take measurements in the whole cross-section during one, single day. We also argued, with reference to the main purpose, that no deep water would normally pass into the Kattegat across the shallows (<10 m) south and southwest of Læsø.

It is possible to obtain an indirect estimate of the flow through Læsø Rende by putting the net outflow across the whole section equal to the long-time average freshwater supply to the Baltic QF = 15 000 m^3/s. This assumption implies a yearly mean outflow of surface

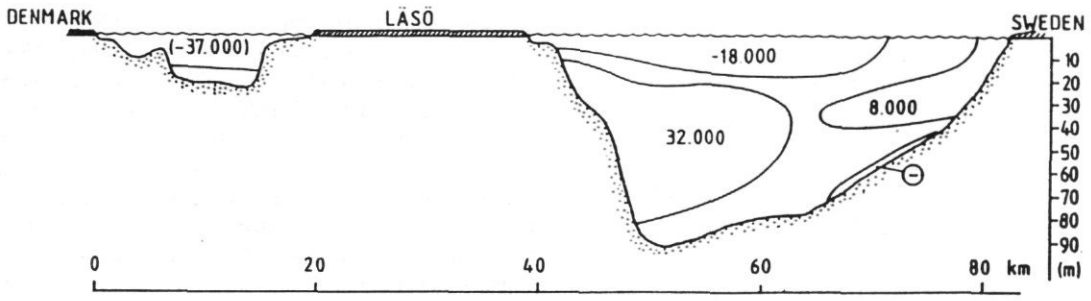

Figure 8. Average transport through the section. Positive values indicate transports into the Kattegat.

water in the Læsø Rende of 37 000 m^3/s, corresponding to a mean velocity northwards in that part of the cross-section of nearly 25 cm/s. According to Dietrich (1951) the observed mean surface velocity at the LV "Læsø Rende" was 22.6 cm/s during the period 1901–1930. In 1912–1913, J. P. Jacobsen carried out intensive current measurements at several depths from the LV "Læsø Rende". Much later, his results were worked up by Rossiter (1968), who determined the mean velocities for a 14-month period of daily measurements. The mean northward velocity is well above 20 cm/s, thus supporting a strong mean outflow of surface water on the Danish side of Læsø.

It is of course tempting at this stage to use old data on salinity and nutrients from Læsø Rende to calculate full cross-section fluxes. This procedure is easily done for salinity. Assuming steady state, the salt flux through the full cross-section must be zero, implying that the salt transport through the Læsø Rende, M_s (Læsø Rende) equals to $M_s(S = 35)$. Daily salinity measurements from

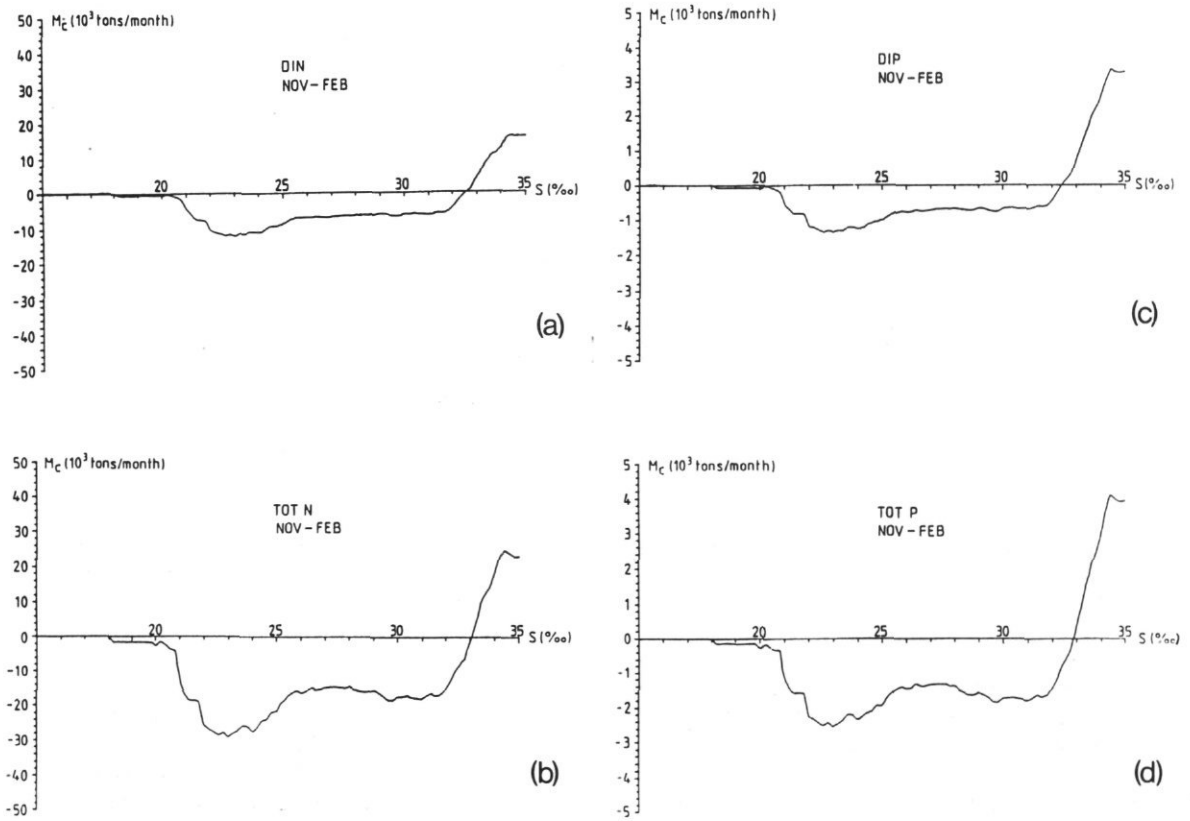

Figure 9. (a) The transport of DIN through the section as a function of salinity. Winter. Positive values indicate transports into the Kattegat. (b) As in 9a but for TOTN. (c) As in 9a but for DIP. (d) As in 9a but for TOTP.

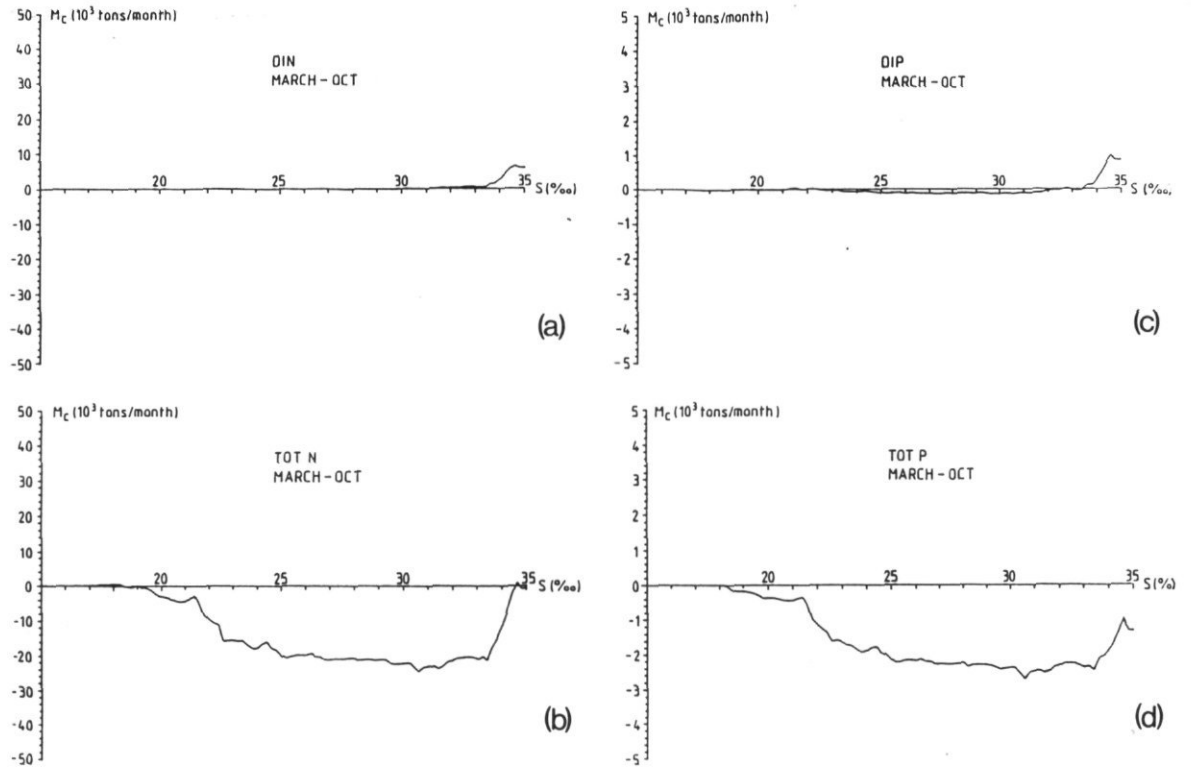

Figure 10. (a) The transport of DIN through the section as a function of salinity. Summer. Positive values indicate transports into the Kattegat. (b) As in 10a but for TOTN. (c) As in 10a but for DIP. (d) As in 10a but for TOTP.

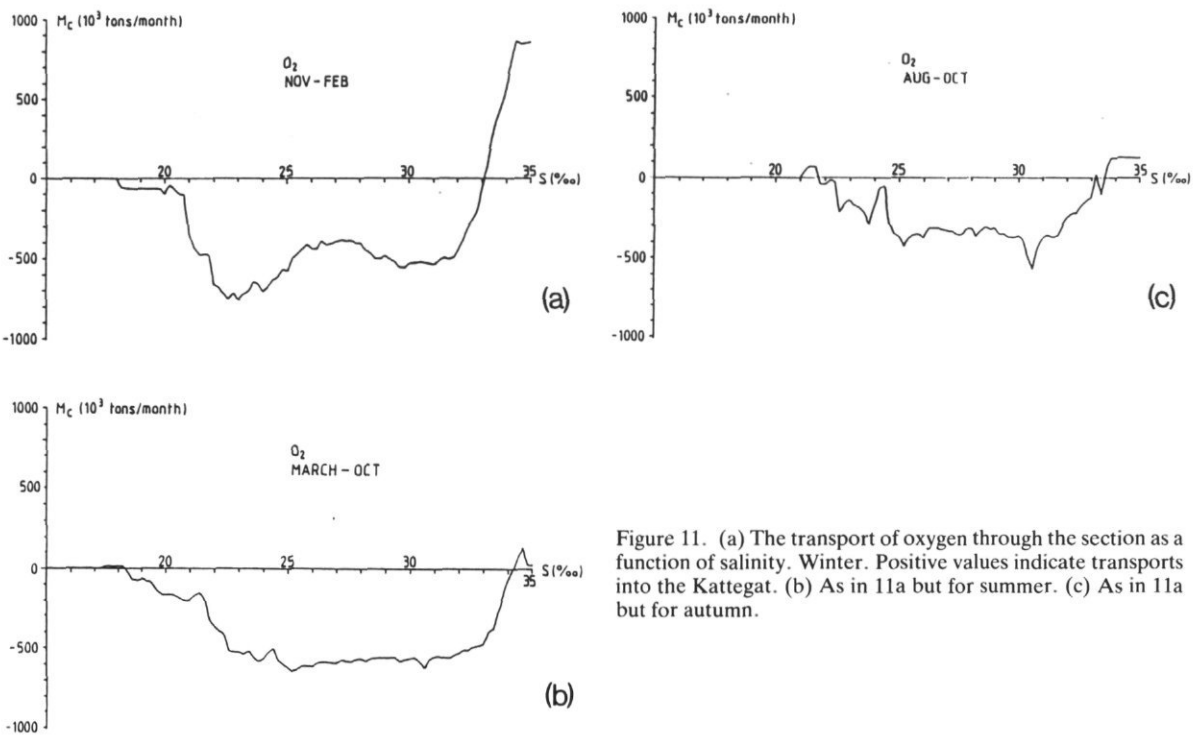

Figure 11. (a) The transport of oxygen through the section as a function of salinity. Winter. Positive values indicate transports into the Kattegat. (b) As in 11a but for summer. (c) As in 11a but for autumn.

several depths exist far back to the beginning of the century. Long-time salinity averages were obtained from the Danish Agency for Environmental Protection (Aertebjerg, pers. comm.). The volume transport M(S) for the full cross-section, including Læsø Rende, is shown in Figure 12. The corresponding salt transport M_s is shown in Figure 13. As seen from this figure, the salt flux for the whole cross-section ends up at exactly zero, which is strong support for the results obtained through this project, even if our calculation on salt flux does not take into consideration that current velocities (direction) and salinities are not quite independent (a feature which is obscured by the averaging process).

Concerning nutrient fluxes, a straightforward calculation is not easily carried out. First, there are relatively few observations. In Table 1, we have summarized Danish and Swedish nutrient and oxygen measurements from Læsø Rende during 1981–1989 (SMHI, database). Flux calculations based on these measurements require a seasonal division of the volume flow through the Læsø Rende, as was done for the eastern part of the section. A seasonal division requires a seasonal estimate of the supply of water from the Baltic. Such estimates have been presented by Svansson (1975), but the variations in atmospheric pressure and sea levels are such that a four-year period, like ours, with relatively few measurements is unlikely to coincide with such long-time averages (different from the freshwater supply in itself).

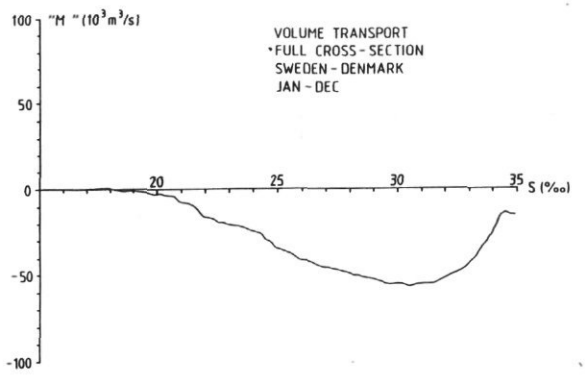

Figure 12. The integrated volume transport through the whole section, including Læsø Rende, as a function of salinity. Positive values indicate transports into the Kattegat.

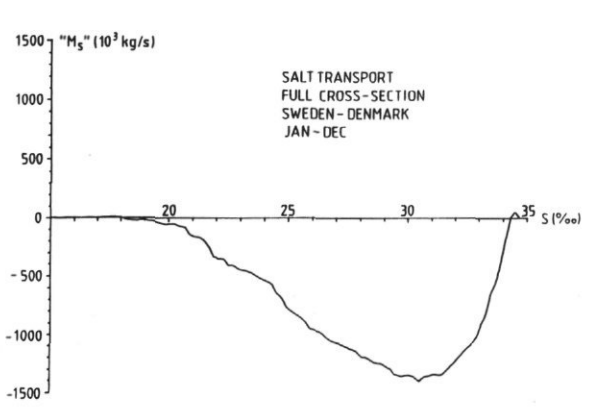

Figure 13. The integrated transport of salt through the whole section, including Læsø Rende, as a function of salinity. Positive values indicate transports into the Kattegat.

Table 1. Salt, oxygen, and nutrient concentrations at Læsø Rende in μmol/l. Surface water S < 30 PSU.

Parameter	Winter	Summer
Salinity	26.19	24.12
Oxygen	7.02	6.81
DIP	0.61	0.22
TOTP	1.09	0.87
DIN	5.16	1.26
TOTN	22.56	18.08

Although one cannot recommend calculation of full cross-section fluxes of nutrients this is no serious drawback. The purpose of this project was to study the deep water fluxes, including the seasonal variations concerning these fluxes. The results obtained here differ significantly from estimates obtained from entrainment theory. An empirical, yearly mean value for the upward entrainment velocity in the Kattegat of 0.25 m/day was derived by Stigebrandt (1983). The measurements presented here indicate a deep-water inflow of 64 000 m^3/s during winter and 28 000 m^3/s during summer compared to roughly 50 000 m^3/s during winter and 20 000 m^3/s during summer for the entrainment approach.

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Interannual variation in wind energy in the Danish Straits

Torben Schelde Jacobsen†

Jacobsen, T. S. 1992. Interannual variation in wind energy in the Danish Straits. – ICES mar. Sci. Symp., 195: 474–477.

Long-term variations in the mixing energy of wind are of interest for studies of hydrographic variations in the Baltic and the oxygen depletion observed in the Danish Sea area in recent years. Preliminary results of an investigation into the effects of wind in the Great Belt are reported on the basis of material from 1891 to 1988, including estimates of mixing energy from histograms collected from Danish lighthouse stations. A comparison between the lighthouse station data and meteorological mast observations for a shorter period is presented and its results discussed.

T. Schelde Jacobsen: Late of the Institute of Hydrodynamics and Hydraulic Engineering, Technical University of Denmark, DK-2800 Lyngby, Denmark.

Introduction

This report gives some preliminary results of an investigation into the long-term variation in the mixing energy of the wind regime in the Great Belt. The investigation is intended to comprise both the current and the wind as sources of mixing, the combined period of investigation being 1891–1988. Up to now, only wind data have been available, and the mixing energy of the wind has been estimated from monthly histograms of wind speed and direction as reported from three lighthouse stations in the Great Belt area.

The histograms were collected from 15 Danish lighthouse stations by Kristensen and Frydendahl (1991) and financed by the National Agency of Environmental Protection in Denmark as part of the HAV90/Marine Research Programme. The present project on mixing energy is financed by the same programme.

A knowledge of long-term variations in mixing energy is of interest in discussion of the variations in the hydrography of the Baltic, and for clarifying the natural component of the driving forces of the oxygen depletion in the Danish Sea area observed in recent years.

Processing of data

The Danish wind climate has been investigated for a 116-year period by means of wind observations from 15 lighthouses. The results are two-way tables (histograms) of the wind in 8 directions and 12 forces (Beaufort scale) on a monthly basis (Kristensen and Frydendahl, 1991). Until recently, the observations were visual and based on a subjective judgement of wind force, but anemometers are now in use at all stations. Although the subjective estimation carries a 'personal factor', it does have certain advantages over the instrumental measurement: the anemometer measures wind speed at a specific point and is very dependent on shadow effects. The subjective estimation reads wind speed from its impact in a larger region: waves at sea, drift of smoke, or movement of branches and trees on land. It is an integral measurement, and an experienced observer may be able to produce reliable results. Care has to be taken when comparing wind energy before and since this change of procedure.

The Beaufort scale in itself also poses a problem. The Commission for Maritime Meteorology under WMO found in 1964 that the WMO Beaufort scale over force 8 gives wind speeds that are too high, and below force 5 the estimate is 10–20% too low. The comparison was based on the effect of the wind on the sea. Therefore, for scientific purposes, a modified Beaufort scale, the 'CMM IV-Scale' was proposed. Although the original Beaufort scale is still in use when reporting wind speeds within the meteorological services, the CMM IV version has been used for the present work in order to obtain better results.

The different ways of estimating/measuring the wind, the reporting in physical units (knots or m/s) or Beaufort, and the subsequent translation present serious obstacles to the use of the data for estimating climatic variations.

The ideal observation would be at sea, well away from

† Deceased 8 October 1992.

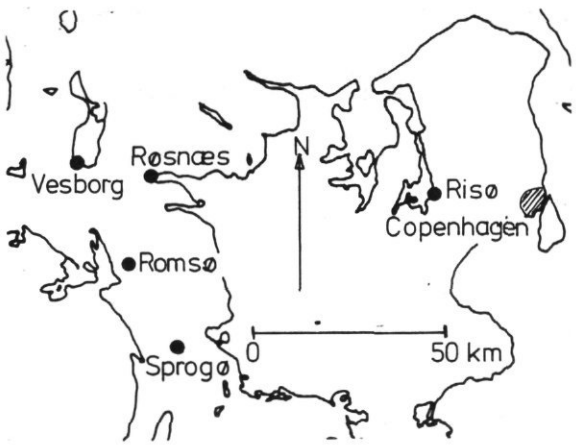

Figure 1. Location of the stations.

the frictional dampening of the wind by roughness elements of the landscape. Since this is impossible we must consider the special features of the coastal stations, and concluding whether or not a long-term change in the wind energy has taken place must be based on coherent results from several stations.

Three stations (Romsø, Røsnæs, and Vesborg, Fig. 1) are available in the Great Belt area, and data have been processed from 1873 to 1988.

As mentioned by Petersen *et al.* (1981) the wind speed distribution can be fitted with a Weibull distribution, which is governed by two parameters: A and c.

$$f(V) = c/A^*(V/A)^{c-1} \exp[-(V/A)^c]$$

c is usually in the range 1.5–2.6, frequently close to 2, which yields a Rayleigh distribution (c = 2). c is a dimensionless shape parameter, while A has a velocity dimension.

The transfer of wind energy to the sea is the product of wind shear stress on the water surface (proportional to V squared) and the surface velocity (usually proportional to V). Thus the rate of energy transfer becomes proportional to V^3. The cubic wind has been used in this report as a measure of the energy transferred to the sea. It is evident that strong winds of even short duration have a high contribution since they are weighted by the cube.

The energy transfer creates turbulence in the water, and in the presence of a two-layer density stratification (which is often found in the Great Belt) the lower layer is eroded (upwards entrainment). The entrainment is associated with an increase in the potential energy of the fluid, since the centre of gravity is raised and the entrainment velocity is directly proportional to the energy input from the wind. This is why the wind cube is an important parameter for studying the mixing conditions.

When the Weibull parameters A and c have been estimated, it is easy to calculate the average wind cube, since

$$\text{Average } (V^m) = A^m \text{Gamma}(1+m/C)$$

and in particular

$$\text{Average } (V^3) = A^3 \text{Gamma}\,(1+3/C)$$

where Gamma is the Gamma function.

A Weibull distribution was fitted to each of the monthly histograms of wind speed and direction, all directions being collected within each wind-speed interval. The observed cumulated probability was next calculated for each interval. It can be shown that the probability of exceeding the velocity V_0 is:

$$P(V > V_0) = \exp(-aV_0^c)$$

$$A = a^{-1/c}$$

and thus one obtains:

$$\ln(-\ln P(V > V_0)) = c \ln V_0 - c \ln A.$$

By choosing V_0 as the series of lower class limits in the CCM IV Beaufort scale, A and c (and a) can be evaluated by linear regression analysis.

Results and discussion

The annual averages of the wind cube are shown in Figure 2 for each of the three stations. The resemblances are small with respect to averages, peak values, and trends. The most spectacular common feature is a rise beginning in the middle 1960s, and ending in a fairly high value in the 1980s. The statistical properties are as follows:

Annual mean m^3/s^3	1873–1988	s.d.
Romsø	559	132
Røsnæs	827	230
Vesborg	748	253

The average wind cube over the sea surface in Denmark is estimated from Petersen *et al.* (1981) at 663 m^3/s^3 at the 10-m level. The data period was 1965–1977 and this long-term average is not expected to show a significant variation over the Danish area.

The three series have certain disadvantages:

- The station surroundings have a varying roughness in different directions, only some of them corresponding to the sea surface roughness (i.e. where the shore is close to the station).

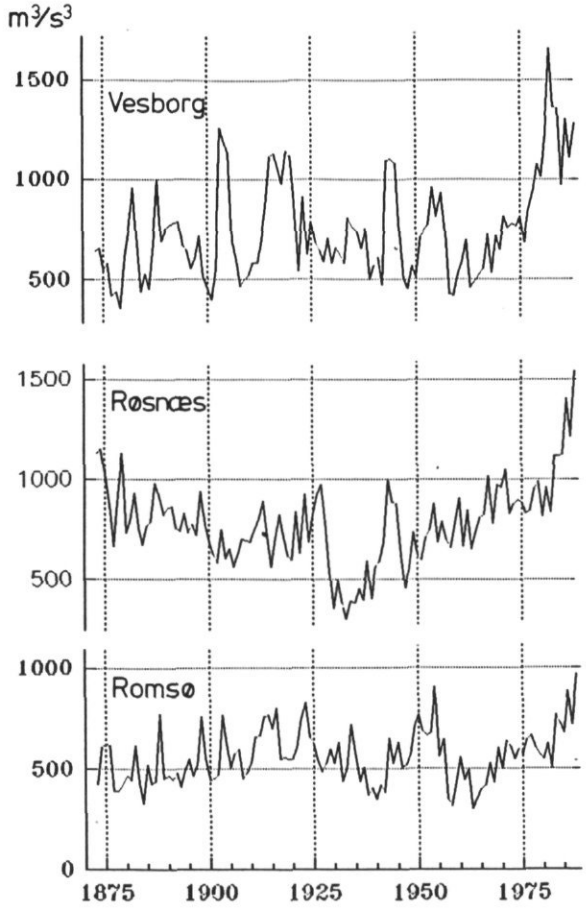

Figure 2. Annual averages of the cubic wind.

- The measuring technique has changed from estimating wind force to reading an instrument.
- Observers have changed over the years.

An obvious question is the extent to which spatial variations can be expected in the mean annual wind cube within the Danish area, given that the "true" wind is known. This can be answered partly by comparing data from the meteorological masts at Sprogø and Risø (70-m and 76-m levels). These two sites have been compared by Abild and Nielsen (1991), who analysed certain features and one result is that the mean wind at Sprogø is reduced by approximately 8% at 70 m, possibly due to a wake effect from the lighthouse hill about 200 m west of the mast. Time overlap of the two series has not yet been analysed in the present investigation, but data from Sprogø reduced to the 10-m level over a sea surface have been available for the period 1980–1987.

The surface roughness distribution differs at the three stations: at Romsø (lowest average and s.d.) there is roughness class 1 ("open areas without significant wind breakers") extending 1.2 km towards the west and somewhat less towards the SW and NW, but an order of magnitude smaller in other directions. The correlation between westerly winds and high wind speeds would indicate a significant reduction in the wind cube. Anemometers have been used since 1985. A few months of data have been substituted from Omø during 1980–1984. Røsnæs has only 10–20 m class 1 roughness between the site and the sea in all directions, but is situated at 13 m height with a strong slope to the sea in all directions except towards E. This indicates possible topographic effects. The wind has been measured with anemometers since 1985. Vesborg has 100 m class 1 towards the SE, S, SW, and W, 60 m towards the NW, and the remaining quadrant between E and N is entirely class 1 (out to a distance where the roughness no longer matters). Anemometers were introduced in 1981.

The difference in average and s.d. is only about 10% for the two last stations, which clearly have a higher energy than Romsø, agreeing qualitatively with the roughness distribution.

The average wind cube at the 76-m level at Risø 1965–1977 (excluding 1964, 1970, 1973, and 1974 because of obvious errors) is 738 m^3/s^3, while the estimate by Lundlaug Petersen *et al.* (1981) at the 10-m level over the sea is 663 m^3/s^3 for the same period. The ratio is 0.90 and the annual Risø mast values have been multiplied with this factor in order to estimate the equivalent wind cube at the 10-m level over the sea. The Risø data are available for 1959–1978 at present, and the rest of the series will be obtained later.

Sprogø mast data have been transformed to the sea at 10 m, but they are only available for 1980–1987 at present. Both are shown as filled circles in Figure 3, since

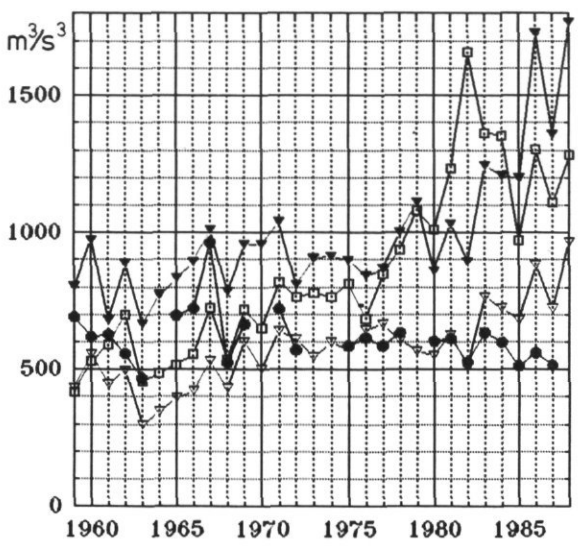

Figure 3. Annual averages of the cubic wind. Sprogø and Risø (filled circles), Romsø (open triangles), Røsnæs (filled triangles), and Vesborg (squares).

they cannot be mistaken for each other. It remains to investigate an overlapping period. At present the filled circles are assumed to represent the best estimate of the annual wind cube average at sea in the Great Belt region. The variance of the Risø data (1959–1978) is markedly higher than that of the Sprogø data (1980–1987), but the average is almost the same. The series of filled circles is now called the 'standard'.

The observations from Røsnæs (filled triangles) are clearly higher than the standard, and the difference escalates from about 1980; 1986–1988 should be treated carefully because of the introduction of the anemometer. Vesborg starts at a lower level than the standard but shows a trend and doubles the values during the 29-year period; 1982 shows a very high peak, but anemometer use began in 1981. Romsø also starts at a lower level than the standard, agrees fairly well in 1968–1982, but then starts climbing, although less so than the other two. An anemometer was installed in 1985, which is when the deviation from the standard is largest.

The preliminary conclusion is that the lighthouse measurements are of dubious quality with the present processing to annual mean wind cubes. Romsø seems to be the most reliable of the three stations investigated. The wind cube during the 1980s is best represented by the Sprogø mast data, the average being 570 m^3/s^3. This is 14% lower than the 1965–1977 value from Petersen *et al.* (1981) of 663 m^3/s^3, but the difference it not necessarily significant: it corresponds to a change of the mean wind of only 2%.

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Hydrographic variability in the Northern Baltic in the twentieth century

Pekka Alenius and Jari Haapala

Alenius, P., and Haapala, J. 1992. Hydrographic variability in the Northern Baltic in the twentieth century. – ICES mar. Sci. Symp., 195: 478–485.

Temperature and salinity variations in the Northern Baltic Sea were studied on the basis of fixed oceanographic stations around the coast of Finland. Data from individual stations were merged by sea areas. The mean sea surface temperature of the Baltic has tended to resemble that of the whole Northern Hemisphere. The recent global trend towards increasing temperature has not, however, been observed in the Baltic area, where the 1930s have been the warmest decade of the century. The 1980s were hydrographically normal, with no indication of an increase in year-to-year variability. During this century salinity has varied by about 1.5 PSU. The salinity minimum prevailed in the 1930s and the maximum in the late 1970s.

Pekka Alenius and Jari Haapala: Finnish Institute of Marine Research, PO Box 33, SF-00931 Helsinki, Finland.

Introduction

The Baltic Sea is a semi-enclosed sea with a positive water balance. Its physical characteristics are determined by energy exchange between the sea and the atmosphere, river runoff, water exchange with the North Sea, and the annual variability of solar radiation. It is a shallow sea, mean depth 60 m, with strong stratification. A permanent halocline prevails at around 80 m depth in the Central Baltic Sea. The seasonal thermocline forms at 10–30 m depth during the summer and northern parts and coasts of the Baltic freeze every winter.

Owing to its physical nature the Baltic reacts strongly to climate change. For example, climate change altering the water exchange and river runoff, and thus modifying the water balance, changes the salinity conditions and stratification. Increasing temperature can also change the stratification of the Baltic.

Data sets

Temperature and salinity of the Baltic have been regularly measured by scientists at the Finnish Institute of Marine Research (FIMR) and their predecessors since 1898. The data sets consist of both open-sea data from research cruises and coastal data from fixed oceanographic stations. Some of these time series have been used earlier in climate studies (Launiainen and Koljonen 1981).

In this study we have used temperature and salinity measurements from the fixed hydrographic stations and lightships. These observations are quite suitable for climate studies. The sampling strategy has remained the same and observation intervals are regular, unlike open-sea observations which are highly irregular both in time and space.

Sea surface temperature (SST) was measured daily and depth observations were carried out on the 1st, 11th, and 21st days of each month throughout the year. The depth observations were made at standard depths of 5-m intervals down to 20-m depth and 10-m intervals below that. The observation dates varied in particular during late autumn and winter, and many observations are missing because of bad weather conditions.

These stations have been based on lighthouses, lightships, coastguard stations, and pilot stations in the outer islands (Fig. 1). The automation of lighthouses and the withdrawal of light vessels have put an end to observations in many cases. Also, the number of pilot stations has been reduced, and this too has led to further decreasing the number of observation stations. In 1910 there were 18 stations and in the 1930s more than 30. At present there are 9 stations, two of them in operation since the beginning of the century (Harmaja and Utö).

Data processing

All the digitized data sets were first screened, and outliers were identified and corrected whenever poss-

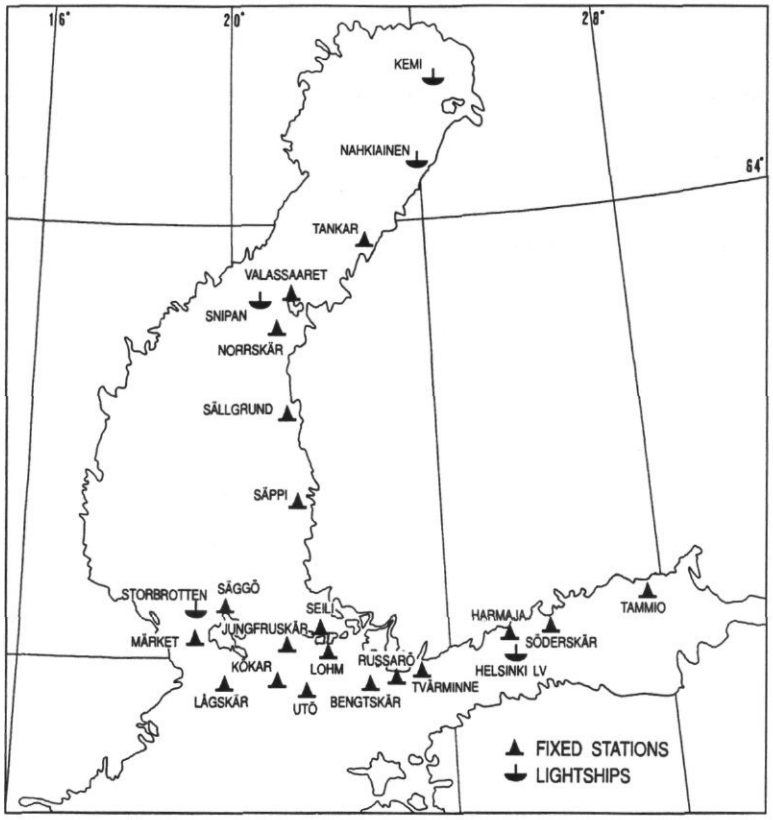

Figure 1. Northern Baltic Sea and the stations used.

ible. The original observation forms were used for corrections. The homogeneity of the time series was checked by comparing the time series with those of two neighbouring stations.

Oceanographic observations are generally difficult and expensive to carry out in the open sea. Reliable automatization of the work has been possible only in the last two decades. Therefore oceanographic time series are much fewer and more sparse than meteorological time series. Long continuous time series exist from only a couple of stations. However, we have tried to synthesize the existing data to determine whole sea areas as follows:

(1) Only those stations in operation for over 10 years between 1951 and 1970 have been considered in the analysis (Fig. 1). This period, chosen in order to have the maximum number of stations in the analysis, has also been used in several other studies (e.g. Jones *et al.*, 1986; Alexandersson and Eriksson, 1989).
(2) The average annual cycle of the temperature and salinity in the reference period 1951–1970 was determined by Fourier analysis for irregularly timed data (Kahma and Voipio, 1990) for all stations and depths.
(3) The mean annual cycle was removed from the time series.
(4) Anomaly time series derived for each station (Fig. 2) were then merged by sea areas (Bothnian Bay, Bothnian Sea, Åland Sea, Archipelago Sea, and Gulf of Finland).
(5) The annual and seasonal averages were computed.

Results and discussion

General remarks

In this study we used a period from May to September to determine year-to-year variability, because most of the stations operate only in the ice-free season and it is not possible to calculate the annual values based on whole-year observations. On the other hand annual means of sea water are weighted values of the thermal summer, because during October–April water temperature is mainly between 0 and 5°C in the Northern Baltic.

The amplitude of year-to-year variability of the SST, about 4°C, seems to have remained fairly constant throughout the century. The changes in temperature and especially in salinity have similar characteristics at all depths and sea areas in the Northern Baltic area (Figs. 4, 5, 6, and 7).

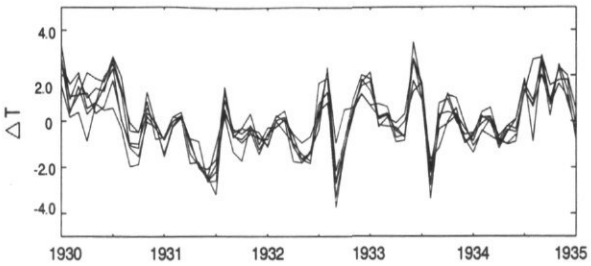

Figure 2. Sea surface temperature anomaly time series from all the stations in the Gulf of Finland. These observations produce the mean temperature of the Gulf of Finland.

The first decades of the century were cool, the surface layer relatively cooler than deeper layers. Considerable warming occurred from the 1920s to the end of the 1930s, when a cooling period until the middle of the 1960s occurred. An exception was the surface layer in the Gulf of Finland. Slight cooling has continued in the Bay of Bothnia and the Archipelago Sea. The near bottom water in the Gulf of Finland and the Archipelago Sea is today even colder than at the beginning of the century.

The SST of the Baltic Sea reflects a tendency similar to the whole northern hemispheric mean (Jones *et al.*, 1986; Houghton *et al.*, 1990). The main difference is in variances (Fig. 3). This is obvious, because hemispheric mean is based on several thousand stations and covers several different climate regions, whereas our data represent local conditions in a limited area. Warming till the 1930s and cooling after that is clear in both time series. Globally, the 1980s have been the warmest decade in this century, but in the Baltic Sea area the 1980s were a very normal decade. Similar results have been presented also in other studies from the Baltic area (Alexandersson and Eriksson, 1989; Astok *et al.*, 1990) and North Atlantic (Antonov, 1990).

In the deepest parts of the Central Baltic salinity variations are controlled mainly by the water exchange between the Baltic and the North Sea. Because of the displacement of the old stagnant water, observed salinity changes are usually rapid and extensive (Fonselius *et al.*, 1984; Matthäus, 1984). Salinity variations in the surface layer reflect changes in the water balance of the Baltic Sea (inflow/outflow, river runoff, precipitation/evaporation). Especially in the Northern Baltic this is a slowly changing parameter and its year-to-year variation is fairly small. The changes, however, are clear on a scale of decades.

During this century salinity has varied ~1.5 PSU. The minimum prevailed in the 1930s and the maximum late in the 1970s. During the first decades of the century salinity was lower than it is nowadays. Salinity began to increase slowly during the 1930s and continued for about two decades. It remained more or less constant until the end of the 1960s before increasing again. During the 1980s the mean salinity of the Baltic decreased (Figs. 5 and 7).

Changes in salinity are first observed in the Gulf of Finland and in the Archipelago Sea. Corresponding changes in the Gulf of Bothnia are seen a couple of years later. For example, a salinity minimum existing in the Gulf of Finland and the Archipelago Sea in 1930 was observed in the Gulf of Bothnia in 1936–1941. The latest strong salinity pulse from the North Sea to the Baltic occurred in 1976 (Astok *et al.*, 1990). Salinity maximum was observed in the Archipelago Sea two years later in 1978, in the Gulf of Finland in 1979, and in the Bay of Bothnia in 1981.

The 1980s compared to other decades

The daily SST observations from Harmaja (32 426 observations) were used in a simple analysis to categorize the 1980s. Decade mean values and standard deviation of SST were computed for this century (Fig. 8). The

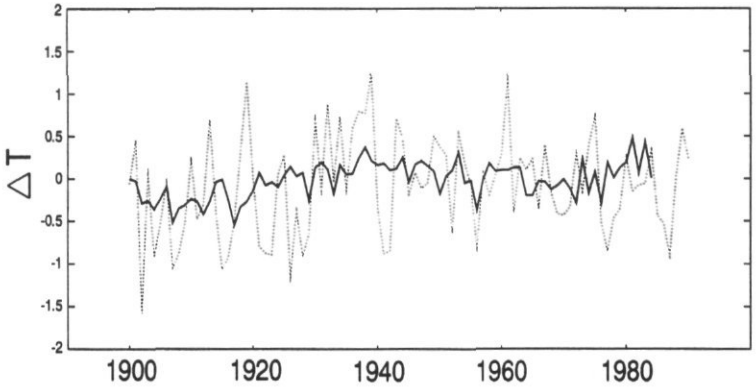

Figure 3. Northern hemisphere mean air temperature anomaly (Jones *et al.*, 1986) and sea surface anomaly from Harmaja lighthouse (dashed). Both time series are unfiltered.

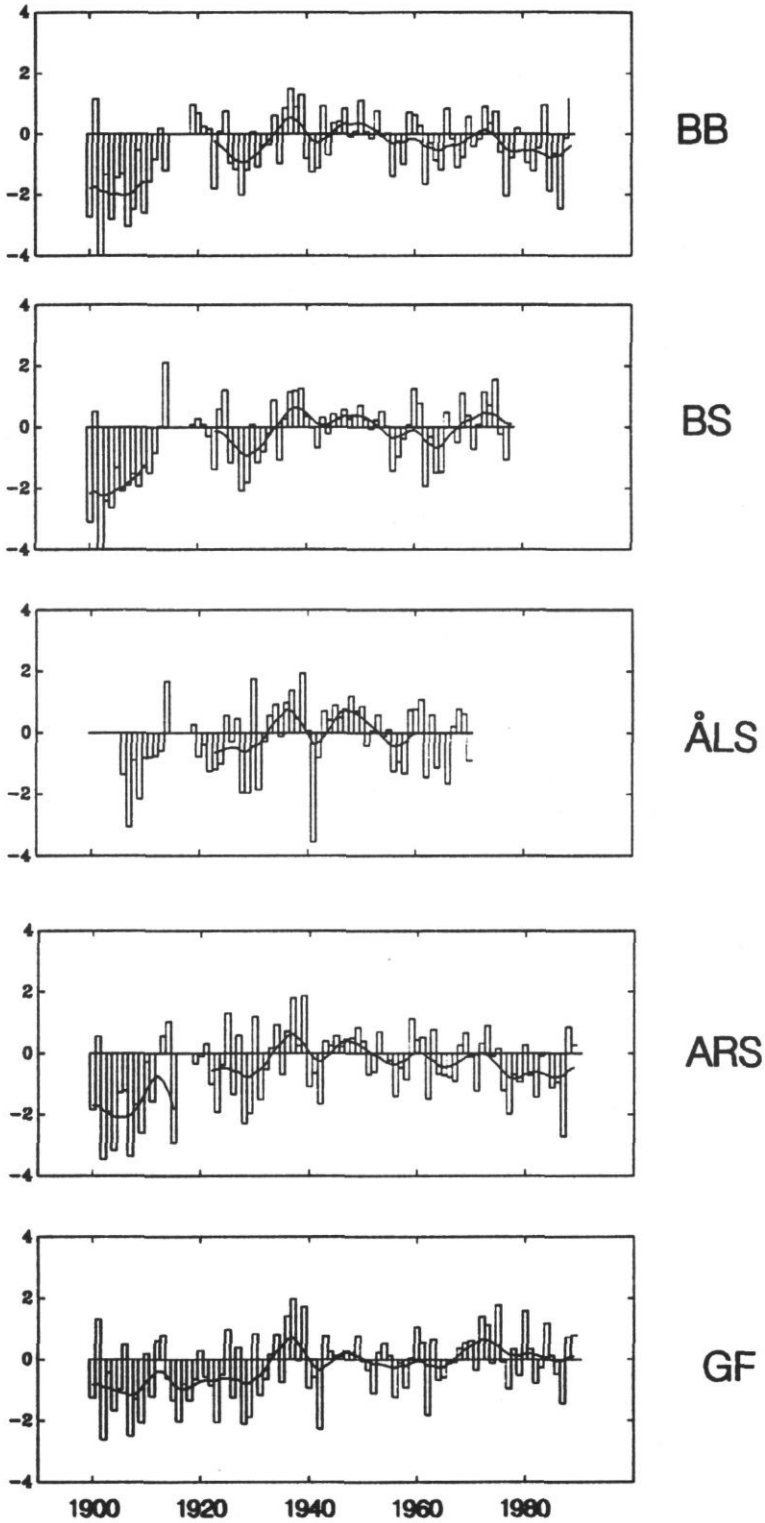

Figure 4. Sea surface temperature anomalies from Bothnian Bay (BB), the Bothnian Sea (BS), the Åland Sea (ÅLS), the Archipelago Sea (ARS), and the Gulf of Finland (GF).

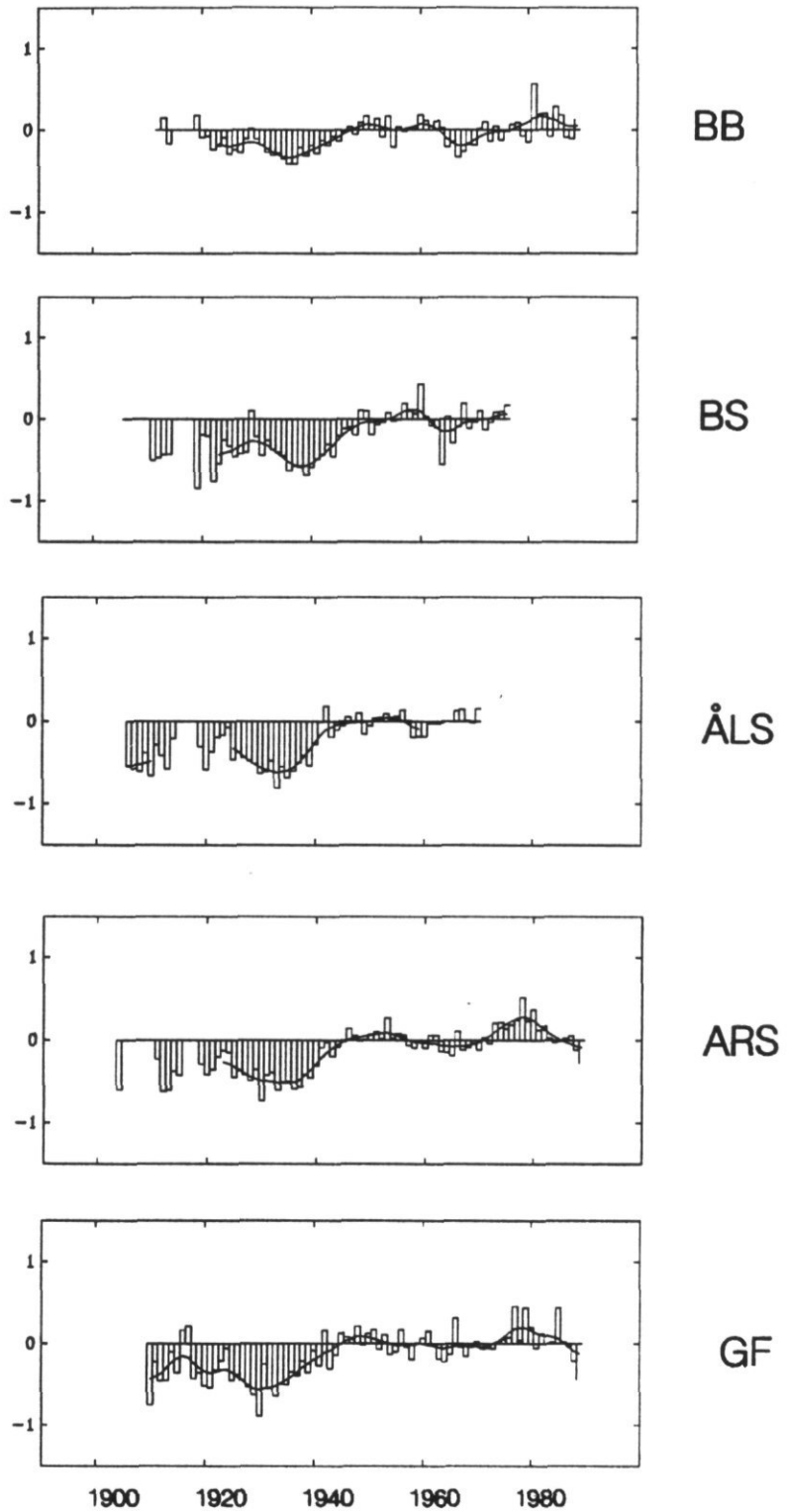

Figure 5. Bottom salinity anomalies from Bothnian Bay (BB), the Bothnian Sea (BS), the Åland Sea (ÅLS), the Archipelago Sea (ARS), and the Gulf of Finland (GF).

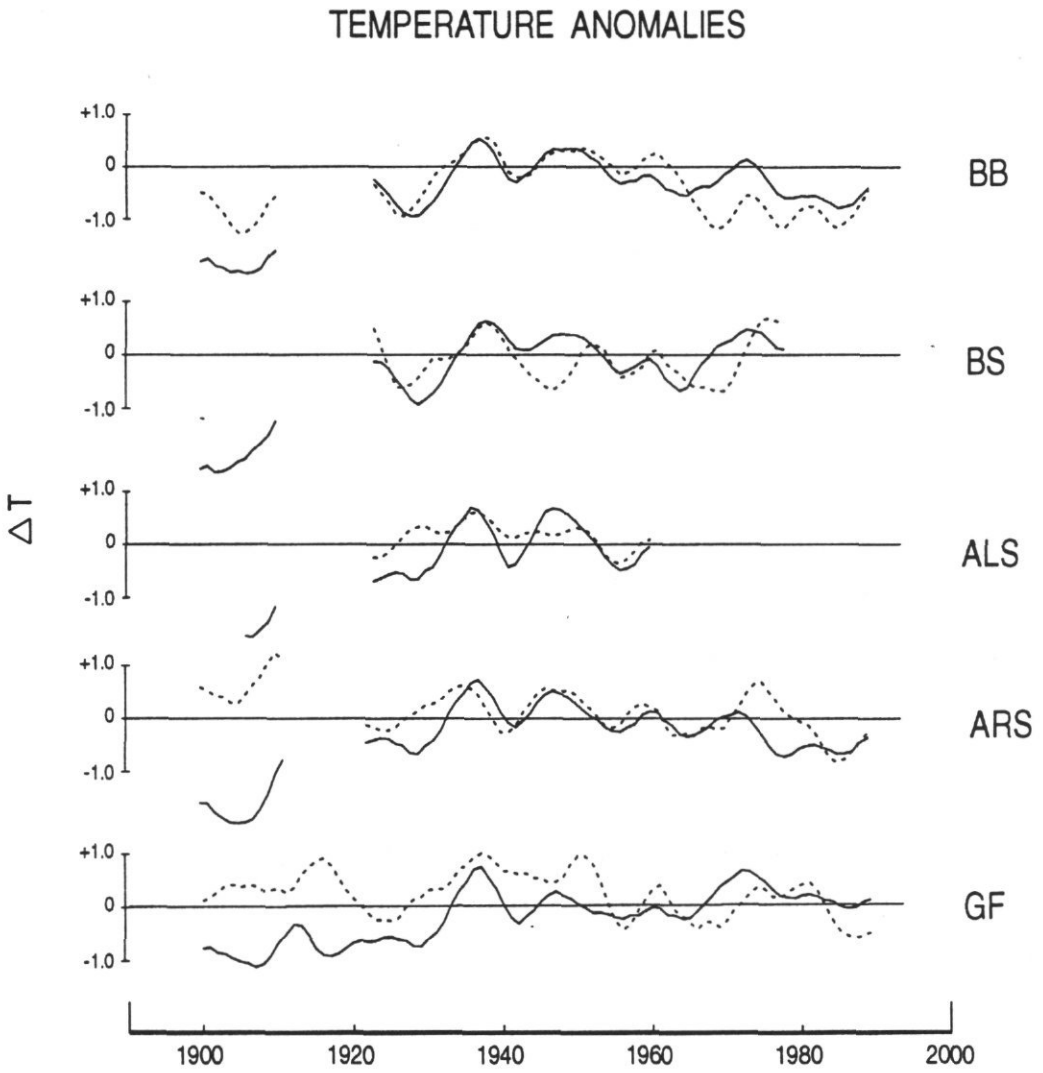

Figure 6. Filtered surface (solid) and bottom (dashed) temperature anomalies.

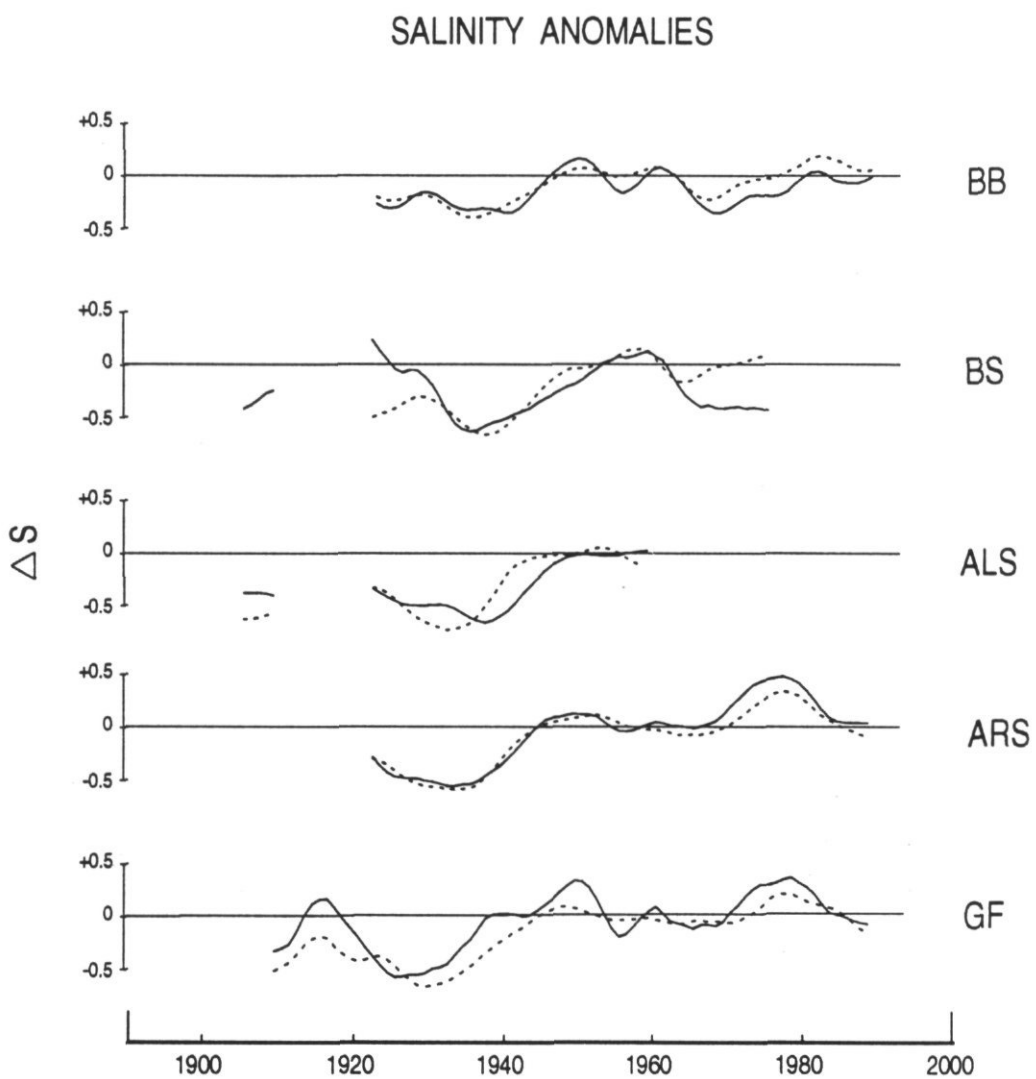

Figure 7. Filtered surface (solid) and bottom (dashed) salinity anomalies.

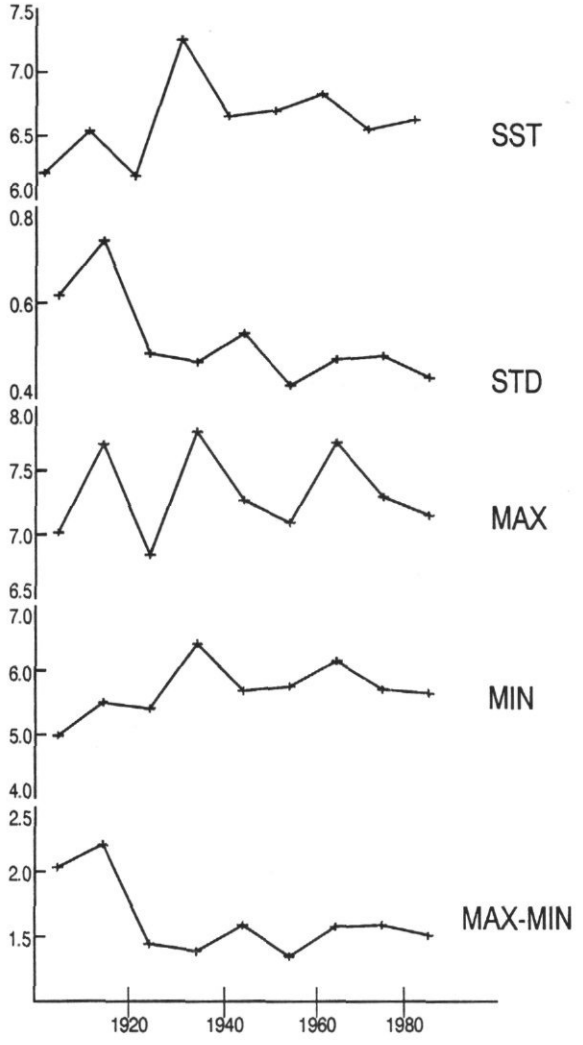

Figure 8. Decade means of sea surface temperature (SST); standard deviation and extremist years of each decade.

warmest and coldest years are also indicated in the figure. The mean temperature of the 1980s was as normal as it could be. The 1980s were the fifth warmest and coldest decade in this century. There has been a lot of debate on the increased climatic variability of the 1980s; this variability has been seen as a signal of changing climate. Our observations, however, do not support this statement. On the contrary, the 1980s were one of the least varying decades in the Northern Baltic Sea. The warmest and coldest years of the 1980s also show that it was a normal decade. During the other decades, both warmer and colder years than the extreme years in 1980s have been observed.

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ICES mar. Sci. Symp., 195: 486. 1992

Long-term trends and variations in hydrographic parameters during the present stagnation period in the deep water of the Central Baltic (Abstract)

Wolfgang Matthäus

Because of major inflows of highly saline and oxygenated water, the Central Baltic deep water was renewed more or less regularly up to the mid-1970s. Stagnation periods lasting up to several years occurred between major inflow events. The present stagnation period, however, differs significantly from those of the past. Lasting more than 14 years in the Eastern Gotland Basin, it has persisted longer than any stagnation period observed during the present century.

Using temperature, salinity, oxygen, and hydrogen sulphide concentration data recorded at the international stations Arkona Deep, Bornholm Deep, and Gotland Deep, long-term trends and variations in selected hydrographic parameters (temperature, salinity, density, oxygen, and hydrogen sulphide concentrations, depths of isohalines and the halocline) have been analysed for various periods and different levels. Special attention is called to the characteristic changes during the present stagnation period in the Central Baltic deep water.

The deep water of the Eastern Gotland Basin has stagnated since 1977. The strongest decreases ever observed were recorded in salinity (−1.7 PSU), temperature (−1.6°C), density (−1.3 σ_t units), and oxygen content (−2.0 cm^3/dm^3; Nov. 1983–1988) at the 100 m level between 1977 and 1988. These resulted in the lowest values in salinity (about 9.1×10^{-3}; 1991) and the temperature (about 3.6°C; 1988) ever recorded at that depth. The decreases in temperature (−1.4 K) and oxygen content (−4.7 cm^3/dm^3) were also extreme at the 200 m level. The salinity of the deep water has been <12 PSU $\times 10^{-3}$ since January 1988. The mean shift of the isohalines to greater depths varied on average between 15 m (9 PSU) and 96 m (12 PSU); the shift of the halocline centre was 9.6 m. During this period, the temperature (7.4°C) and hydrogen sulphide concentration (>5 mg/dm^3) reached the highest values ever recorded in the Baltic sea.

The current stagnation period has caused the most extreme changes in the deep water of the Eastern Gotland Basin ever observed since oceanographic observations started in the Baltic. It must be regarded as one of the longest and most severe stagnation intervals recorded during the present century.

Detailed results of the investigations are published in *Fischerei-Forschung, Rostock*, Vol. 28, No. 3 (1990), pp. 25–34.

Wolfgang Matthäus: Baltic Sea Research Institute, An der Jägerbäk 2, 2510 Rostock-Warnemünde, Germany.

ICES mar. Sci. Symp., 195: 487. 1992

Is the reproduction of Baltic cod governed by oceanographic factors? (Abstract)

S. Carlberg and B. Sjöberg

Cod depend on pelagic eggs for reproduction. In the Baltic, with its low saline water and variable oxygen conditions, this is of particular concern. The cod egg needs sea water with a salinity of at least 10 to 12 PSU and a concentration of dissolved oxygen of at least 1–2 ml/l to develop normally. Furthermore, reproduction of cod is known to take place mainly during the period March–May and only in limited areas in the Baltic. Defined in this way the water volume in which the cod egg can develop until hatching is very limited. In this paper this 'reproduction volume' (RV) is calculated for the reproductive period during some decades and its variation in time is studied. An attempt is made to correlate with known data on variability in the strength of year classes of Baltic cod as defined by data on year class 2 from virtual population analysis (VPA). The results show a positive correlation, albeit weak, between the time series of RV and VPA. The highest correlation coefficients are obtained for the months December–March and with a time lag of nearly two years. This coincides with the spawning period, i.e. the period during which the hydrographical conditions supposedly invoke the largest influence on the reproduction. The results also suggest that during the last decades the salinity and oxygen regime of the Baltic has been sufficient to support a successful cod reproduction and that, therefore, other biotic or abiotic factors may have had a stronger influence on the recent decline in stocks of Baltic cod. Without knowing how great the influences these factors have had on the success of the reproduction, it is not possible to assess the importance of the hydrographical conditions during spawning.

(Editors' note: Since this is a preliminary paper the authors have chosen to continue their work for publication at a later stage).

This is a discussion paper and status report of an ongoing study.

S. Carlberg and B. Sjöberg: Swedish Meteorological and Hydrological Institute, Oceanographical Laboratory, PO Box 2212, S-403 14 Gothenburg, Sweden.

ICES mar. Sci. Symp., 195: 488–491. 1992

The effect of cod predation on the weight-at-age of herring in the Baltic

Henrik Sparholt and Ingvar Bundgaard Jensen

Sparholt, H., and Jensen, I. B. 1992. The effect of cod predation on the weight-at-age of herring in the Baltic. – ICES mar. Sci. Symp., 195: 488–491.

Because cod prey mostly on small food items the small specimens of a herring age group will suffer higher mortality than the large ones. When the predation pressure varies owing to fluctuating amount of cod the mean weight-at-age of herring will change accordingly, even when the individual growth of herring remains unchanged. Based on predation mortalities calculated in the Multispecies Virtual Population Analysis for the Central Baltic the size-specific predation mortality is presented. By using this size-specific predation mortality and a predation pressure proportional to the cod stock biomass in this area in the period 1982 to 1988, it was found that the fluctuating predation results in a 20% variation in the weight-at-age of the herring stock in Subdivisions 25–27. This accounts for a significant part of the variation observed in the weight-at-age in the herring stock in the period investigated.

Henrik Sparholt[1] and Ingvar Bundgaard Jensen: Danish Institute for Fisheries and Marine Research, Charlottenlund Castle, DK-2990 Charlottenlund, Denmark. [1]*Present address: ICES, Palægade 2–4, DK-1261 Copenhagen K, Denmark.*

Introduction

The weight-at-age of herring in the Baltic Sea has varied significantly in the last decade, being up to 30% lower in recent years than at the beginning of the 1980s. This has been observed in commercial catches (Anon., 1991) and in scientific surveys in the feeding areas (Anon., 1990a), as well as in the spawning areas for herring (Anon., 1991).

Several explanations have been given for this variation. One is that it is due to changed migration patterns, with more small herring coming from the northern to the southern parts of the Baltic (Anon., 1988). As growth is faster in the southern part this should increase the growth of the northern herring. This, however, has not been observed on the spawning grounds of the northern stocks. Another explanation is that the cold winters in 1985/1986, 1986/1987, and 1987/1988 would have decreased the growth of herring (Wyszynski, 1988). However, the mild winters in 1988/1989 and in 1989/1990 have not improved growth. A third explanation assumes that the number of food items for herring has been low in later years. This hypothesis has not been examined and it is not certain whether information exists to do so (Teit Reid, pers. comm.). A fourth one, dealt with here, follows the idea by Sparholt (1991) that the variation in predation pressure could explain part of the observed change in weight-at-age. Because the predation mortality is larger on small herring than on larger herring a change in predation stock size would affect the small herring most. Therefore, the change in survival will differ between small and large specimens within an age group and will result in fluctuations in mean weight-at-age. Sparholt found a positive correlation between cod biomass and herring weight-at-age.

Material

Length distribution by age of herring was taken from Sparholt (1990). Table 1 gives the mean length-at-age by quarter for herring in Subdivision 26 in 1985–1989 based on commercial Polish catch samples and from acoustic surveys. The standard deviation of length by age is also given by Sparholt to be: s.d. (cm) = 0.17 + 0.10* length (cm).

Predation mortalities of herring due to cod predation were taken from a Multispecies Virtual Population Analysis (MSVPA) run similar to the one presented in Anon. (1990b), except that the terminal F for sprat in Subdivision 25 in 1985 was changed to a more realistic level, and the biomass of other food from 3 million tonnes to 1 million tonnes.

Table 1. Mean length-at-age for herring in Subdivision 26 in 1985–1986.

Age	Quarter	Length at mid-quarter in cm (l_m)	Length increment in cm (dl_m)
0	3	8.0	
	4	10.8	2.8
1	1	12.6	1.8
	2	14.1	1.5
	3	15.2	1.1
	4	16.1	0.9
2	1	16.8	0.7
	2	17.4	0.6
	3	18.0	0.6
	4	18.6	0.6
3	1	19.1	0.5
	2	19.6	0.5
	3	20.0	0.4
	4	20.4	0.4
4	1	20.8	0.4
	2	21.2	0.4
	3	21.6	0.4
	4	21.9	0.3
5	1	22.2	0.3
	2	22.5	0.3
	3	22.8	0.3
	4	23.0	0.2

Method

It is assumed that the length of herring of age 0 in the stock in Subdivisions 25–27 in the middle of the third quarter is 8.0 cm in all years considered, i.e. 1982 to 1988. It is further assumed that the mean length increments of individual fish are as given in Table 1. Because the s.d. of the length increases with length, it is likely that the larger fish grow faster than the smaller ones within an age group. This is modelled by correcting the mean length increment by the expression:

$$l(a+1) = l_m(a+1) + [l(a) - l_m(a)] * [0.17 + 0.1 * l_m(a+1)]/[0.17 + 0.1 * l_m(a)]$$

where l_m is the mean length from Table 1.

Figure 1 shows a plot of log(M2) against log(length) for all four species for the Central Baltic MSVPA.

The size-dependent predation mortalities, M2s, for each of the four prey species are obtained from an ANOCO with species as class variable, log(length) as the continuous variable, and log(M2) as the dependent variable. Table 2 gives the results of the ANOCO.

Table 3 illustrates the applied procedure. The starting

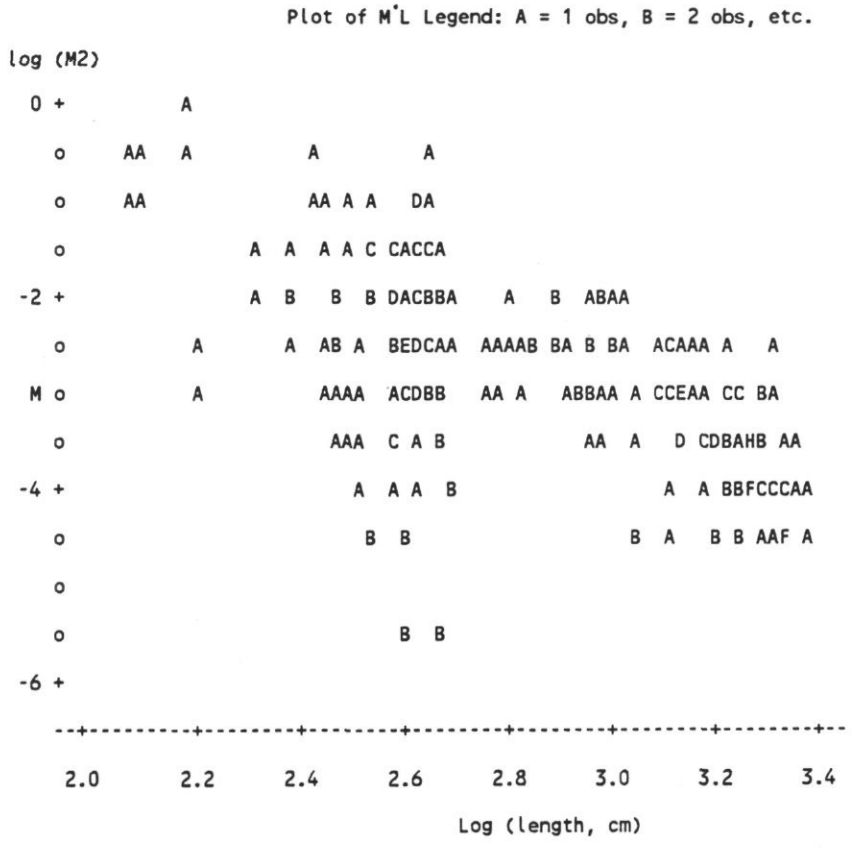

Figure 1. Plot of log (M2) against log (length, cm) for sprat and herring in the Central Baltic. A represents one observation, B two observations, etc.

Table 2. The ANOCO for size-dependent predation mortality.

Error source	d.f.	S. Sq.	Mean S.	F	PR > F	R-square
Model	4	110.01	27.50	41.23	0.0001	0.40
Error	251	167.40	0.667			
Corrected total	225	277.41				

Source	d.f.	S. Sq.	F	PR > F
Species	3	10.78	5.72	0.0008
L	1	61.84	61.84	0.0001

Parameters		
Intercept		5.3007
Species	herring 25–27	0.6009
	herring 28–29s	0.5001
	sprat 25	−0.2431
	sprat 26+28	0.0000
Length		−3.0084

Table 3. The number by length of a cohort at 11 successive quarters, starting at 15 August. Total mortality set to be equal to predation mortality in 1988 (see Table 4).

L	NO	M	NO	L	M	NO	L	M
4.5	14	1.15197	7.87	6.2897	0.43465	5.10	7.4402	0.29370
5.5	261	0.67863	185.90	7.5784	0.26469	142.67	8.9144	0.17851
6.5	1 846	0.43278	1 486.81	8.8670	0.17293	1 250.70	10.3887	0.11646
7.5	4 888	0.29265	4 222.63	10.1557	0.11911	3 748.47	11.8629	0.08013
8.5	4 488	0.20702	4 407.35	11.4443	0.08549	4 046.24	13.3371	0.05746
9.5	1 846	0.15177	1 711.10	12.7330	0.06341	1 605.96	14.8113	0.04259
10.5	261	0.11455	246.47	14.0216	0.04832	234.84	16.2856	0.03244
11.5	14	0.08856	13.39	15.3103	0.03766	12.90	17.7598	0.02527

NO	L	M	NO	L	M	NO	L
3.80	8.3990	0.22498	3.03	9.1021	0.18490	2.52	9.6773
119.34	10.0278	0.13604	104.16	10.8443	0.11170	93.16	11.5124
1 113.20	11.6567	0.08842	1 019.00	12.5866	0.07254	947.70	13.3474
3 459.84	13.2856	0.06066	3 256.21	14.3289	0.04974	3 098.20	15.1825
3 820.31	14.9144	0.04340	3 658.06	16.0711	0.03557	3 557.22	17.0175
1 539.00	16.5433	0.03211	1 490.37	17.8134	0.02631	1 451.67	18.8526
227.35	18.1722	0.02442	221.86	19.5557	0.02000	217.47	20.6876
12.58	19.8010	0.01900	12.34	21.2979	0.01556	12.15	22.5227

M	NO	L	M	NO	L	M	NO
0.15977	2.15	10.0608	0.14480	1.86	10.5082	0.12750	1.64
0.09637	84.60	11.9577	0.08732	77.52	12.4773	0.07706	71.77
0.06252	890.26	13.8546	0.05664	841.24	14.4464	0.05007	800.16
0.04283	2 968.31	15.7515	0.03880	2 855.36	16.4155	0.03434	2 758.96
0.03061	3 423.80	17.6485	0.02772	3 330.18	18.3845	0.02457	3 249.37
0.02263	1 419.19	19.5454	0.02049	1 390.41	20.3536	0.01817	1 365.37
0.01719	213.76	21.4423	0.01557	210.46	22.3227	0.01382	207.57
0.01337	11.99	23.3392	0.01211	11.84	24.2918	0.01075	11.72

L	M	NO	L	M	NO	L
0.8918	0.11481	1.46	11.2753	0.10510	1.31	11.6588
12.9227	0.06952	66.95	13.3680	0.06362	62.83	13.8134
14.9536	0.04523	764.77	15.4608	0.04138	733.77	15.9680
16.9845	0.03105	2 674.61	17.5536	0.02841	2 599.70	18.1227
19.0155	0.02223	3 177.93	19.6464	0.02033	3 113.96	20.2773
21.0464	0.01646	1 343.08	21.7392	0.01505	1 323.02	22.4320
23.0773	0.01252	204.99	23.8320	0.01145	202.66	24.5866
25.1082	0.00974	11.60	25.9247	0.00891	11.50	26.7412

Table 4. Calculated length-at-age (cm) of a herring cohort from age 0 to age 3 at various predator levels, corresponding to biomasses of cod observed in the Baltic in 1982–1988 estimated from bottom-trawl surveys (Sparholt, 1990).

		Cod biomass (million tonnes)						
		0.85	1.27	0.82	0.53	0.26	0.30	0.34
Age	Quarter	1982	1983	1984	1985	1986	1987	1988
0	3	8.0	8.0	8.0	8.0	8.0	8.0	8.0
	4	11.0	11.1	11.0	10.9	10.8	10.8	10.9
1	1	12.9	13.1	12.9	12.8	12.7	12.7	12.7
	2	14.6	14.8	14.6	14.4	14.2	14.2	14.2
	3	15.8	16.0	15.8	15.5	15.3	15.4	15.4
	4	16.8	17.1	16.7	16.5	16.2	16.3	16.3
2	1	17.4	17.8	17.4	17.2	16.9	16.9	16.9
	2	18.2	18.6	18.2	17.9	17.6	17.6	17.7
	3	18.9	19.2	18.9	18.6	18.2	18.3	18.3
	4	19.5	19.9	19.5	19.2	18.8	18.9	18.9
3	1	20.2	20.6	20.2	19.8	19.5	19.5	19.6
		Weight (g) at age 3 during the first quarter						
		59.6	62.7	59.2	56.5	53.7	54.1	54.5

point is the numbers by cm-group of 0-ringers on 15 August of a simulated cohort. A particular M2 value corresponds to each length and the procedure is then to follow each length group from 15 August to 15 February 2.5 years later (i.e. as 3-ringers) as regards its growth and survival. When the average lengths and survival numbers on 15 February are known for each starting length group the average size of the cohort at that time is readily calculated.

Results

Table 4 shows the mean length-at-age of a cohort from age 0 on 15 August to age 3 on 15 February for various levels of predation pressure, corresponding to cod biomass fluctuations in 1982–1988.

Discussion

If the distribution of the length of age 0 on 15 August is more "flat-topped", the size effect will be even greater. If the distribution shown in Table 2 is substituted by, for instance:

5.5 cm	261
6.5 cm	2 846
7.5 cm	3 888
8.5 cm	3 888
9.5 cm	2 846
10.5 cm	261
11.5 cm	14

then the difference in weight between 1983 and 1988 would have increased from 17% to 23%.

In the North Sea, too, it has been shown that the mean weight-at-age of prey found in predator stomachs is less than the mean weight-at-age in the sea. This is not taken into account in the present version of the North Sea MSVPA, but could be incorporated by modifying the weight-at-age along the lines indicated here. Another solution is of course to create a length-based MSVPA.

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ICES mar. Sci. Symp., 195: 492–498. 1992

Possible interactions between salmon migrations and landings, smolt production, herring abundance, and hydrographic factors in the Gulf of Bothnia, 1976–1990

Erkki Ikonen and Raimo Parmanne

Ikonen, E., and Parmanne, R. 1992. Possible interactions between salmon migrations and landings, smolt production, herring abundance, and hydrographic factors in the Gulf of Bothnia, 1976–1990. – ICES mar. Sci. Symp., 195: 492–498.

The greater part of the natural and hatchery production of salmon smolts in the Baltic Sea occurs in the northern part of the Gulf of Bothnia, but the main feeding areas are situated in the surroundings of the Gotland and Bornholm Deeps. The post-smolt run from the home river to the Baltic Main Basin lasts from May/early June to December. According to catch statistics and reported tag recoveries, a varying proportion of post-smolts end their feeding migration in the southern part of the Gulf of Bothnia. One of the most important food items for salmon is the herring. The possible dependence of the salmon catch on the hydrographic factors, herring biomass, and herring year-class strength is considered. According to the analyses the factors best explaining the yearly variation of the salmon offshore catches in the Bothnian Sea were the herring biomass and the smolt production of the previous year.

Erkki Ikonen and Raimo Parmanne: Finnish Game and Fisheries Research Institute, Fisheries Division, PO Box 202, SF-00151 Helsinki, Finland.

Introduction

Baltic salmon originating from the Gulf of Bothnia feed chiefly in the Baltic Main Basin. The main feeding grounds are situated in the area of the Gotland Deep and the Bornholm Deep. However, a proportion of the feeding salmon seems to stay and feed in the Bothnian Sea. This proportion varies widely, forming between 2% and 20% of the total offshore catches. Clupeids constitute the main food of salmon in the Baltic Sea. In the Baltic Main Basin sprat dominate in the salmon diet (Christensen, 1961; Thurow, 1966; Christensen and Larsson, 1979).

In the northern part of the Baltic Sea, herring are more important (Christensen and Larsson, 1979; Linko *et al.*, 1979; Andersson, 1980; Salmi and Ikonen, 1982). The effect of the environmental conditions on the feeding salmon is poorly known. Reddin (1986) has shown that salmon feeding in the Northern Atlantic prefer a temperature range of 4–8°C during feeding migration. The aim of this paper is to find a possible reason for the greatly varying choice of feeding ground and to determine whether the reasons are biological or hydrological.

Salmon in the Bothnian Sea

Salmon smolt production in the Gulf of Bothnia

The salmon fisheries in the Bothnian Sea (Fig. 1) are based on the wild and hatchery smolt production in Subdivisions 30–31 (Fig. 2). In 1989, the contribution of natural production in the Gulf of Bothnia was 0.3 million smolts, whereas that of hatchery production was 3.7 million (Anon., 1991a). The greater part of the natural production came from Swedish rivers. The hatchery production was about equal on both sides of the Gulf of Bothnia (Anon., 1980, 1984, 1987, 1990, 1991a). The Swedish smolt production has been presented as the total figure for the whole Baltic Sea. The annual production in the Main Basin area has been estimated at 300 000 smolts, and this number has been deducted from the total figure.

The most important feeding grounds of the salmon originating from the rivers emptying to the Gulf of Bothnia are situated in the Baltic Main Basin. Of the tag recoveries of hatchery-reared smolts released in 1980–1989 in Subdivision 31 in Finland, 62% were obtained

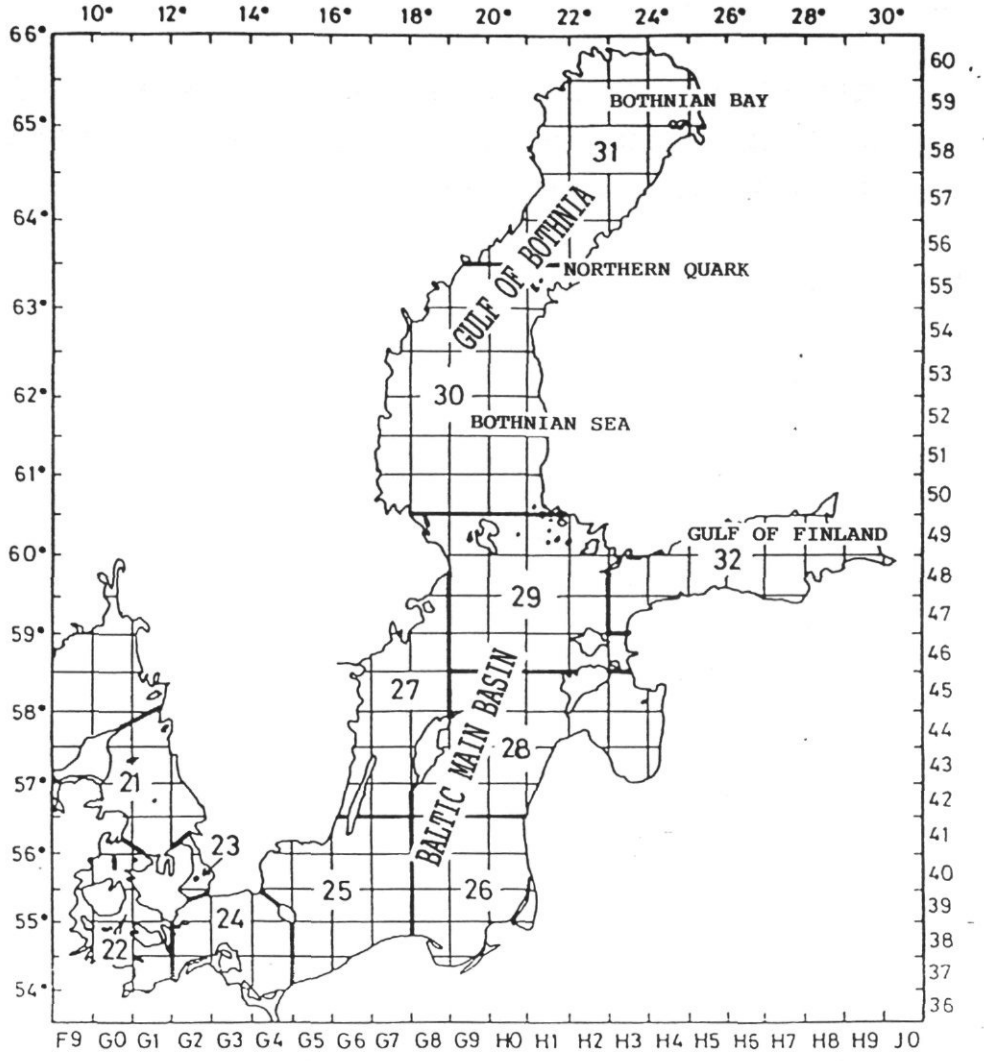

Figure 1. The Baltic Sea.

from the Main Basin, 12% from the Bothnian Sea, 25% from the Bothnian Bay, and 1% from the Gulf of Finland. In Swedish tagging experiments carried out in Subdivisions 30 and 31 (Anon., 1991a) in the same period, the distribution was: Main Basin 85%, Bothnian Sea 11%, and Bothnian Bay 4%; recoveries from the Gulf of Finland were negligible.

The Finnish salmon releases in the Bothnian Sea consisted of River Neva stock, which does not migrate so far. The proportion of recoveries made in Subdivision 30 was 87% (Anon., 1991a).

Salmon fisheries in the Bothnian Sea

Salmon are caught during the feeding migration with drift nets and drifting long lines. The salmon offshore fishery is permitted from the middle of September to the middle of June. A small number of standing lines have been used near the coast in spring. During the spawning run in spring and summer, trapnets have been used almost exclusively.

In 1976–1990, the offshore catches varied between 66 and 544 tonnes (Fig. 3). However, the fluctuation of these landings does not show the same pattern as the offshore catches in the Main Basin or the total catches in Subdivisions 24–31 (Fig. 3). The great majority of the feeding salmon have been caught during September–December with drift nets and long lines (Fig. 4).

Material

Yearly data concerning salmon and herring in the Gulf of Bothnia were available for the period 1976–1990. The following variables were investigated: (1) Salmon catch

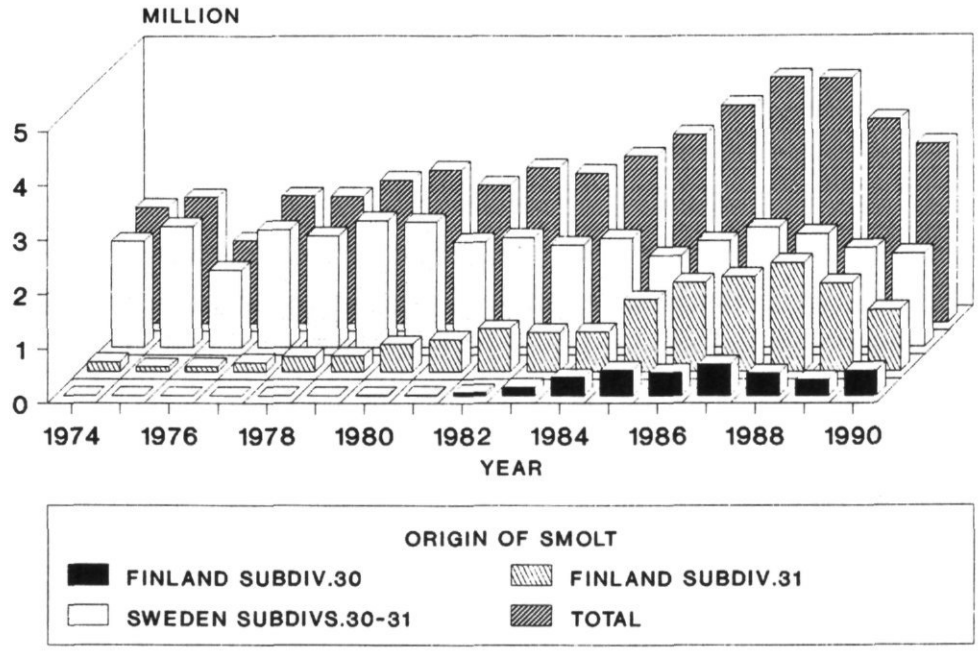

Figure 2. Wild and hatchery-reared salmon smolt production in the Gulf of Bothnia.

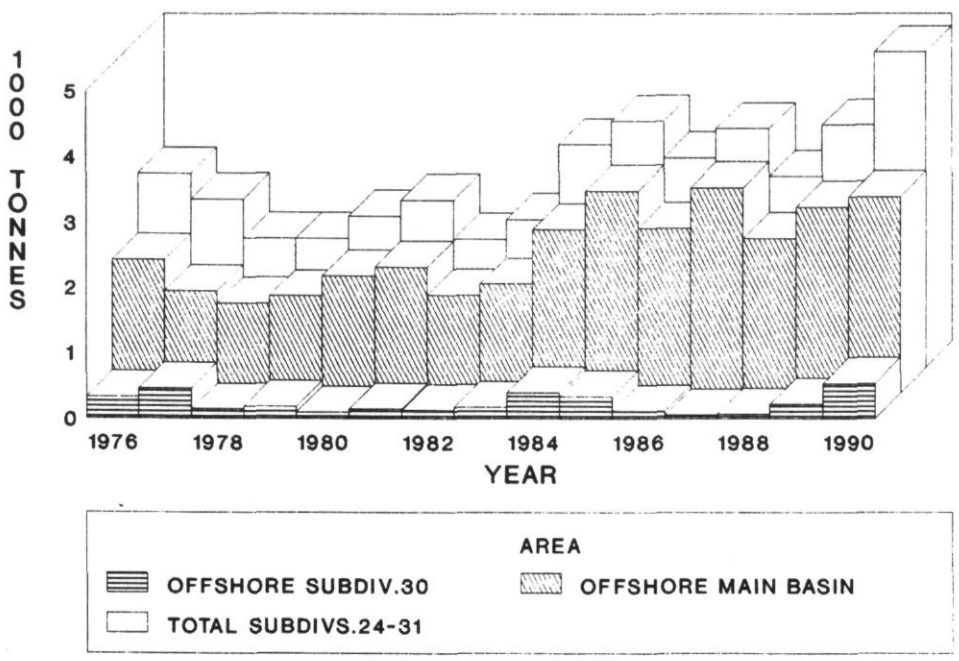

Figure 3. Salmon landings in the Baltic Sea in 1976–1990.

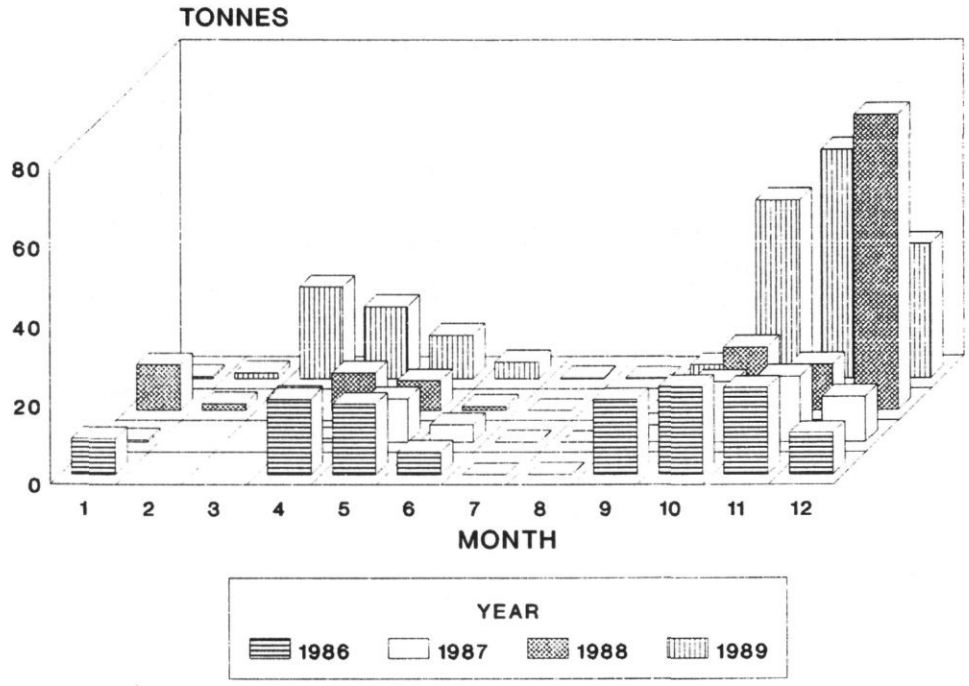

Figure 4. Timing of the landings in the Bothnian Sea offshore fishery.

in the offshore fishery in the Bothnian Sea (ICES Subdivision 30) (Fig. 3). (2) Salmon smolt production of the previous year in the Gulf of Bothnia (ICES Subdivisions 30 and 31) (Fig. 2). (3) Herring year-class strength in the Gulf of Bothnia (Anon., 1991b). (4) Number of one-year-old herring in the Gulf of Bothnia (Anon., 1991b). (5) Herring stock biomass in the Gulf of Bothnia (Anon., 1991b).

In addition, the severity of the winter was studied on the basis of: (6) The number of real ice days in the northern part of Subdivision 30 at Norrskär (Kalliosaari, 1978, 1982; Kalliosaari and Seinä, 1987; Kalliosaari, pers. comm.).

Yearly water temperature records for 1974–1989 from the various depths were obtained from the Finnish Institute of Marine Research as follows: (1) Mean monthly water temperature in July in the northernmost part of Subdivision 30, at Valassaaret, from depths of 0 m, 5 m, 10 m, 15 m, and 20 m. (2) Mean monthly water temperature in October in the northernmost part of Subdivision 30, at Valassaaret, from depths of 0 m, 5 m, 10 m, 15 m, and 20 m.

Methods

The possible dependence of salmon offshore catches in the Bothnian Sea (ICES Subdivision 30) on the herring abundance and hydrographical factors was investigated by calculating the Pearson product-moment correlations (SAS Institute Inc. 1985).

The dependence of the salmon offshore catch in the Bothnian Sea on the smolt production of the previous year, Baltic herring abundance, severity of the winter, water temperature in October of the actual year and in July at the previous year was investigated by multiple regression analysis (SAS Institute Inc. 1987). In the procedure the linear regression model is fitted by the least squares method. The ability of models including various independent variables to predict the salmon catch was studied with the coefficient of determination (R^2).

As the regressor variables in the linear regression model seem to correlate, a principal component analysis (SAS Institute Inc. 1987) was made in order to obtain uncorrelated independent variables for the model. The principal components were computed from the correlation matrix. The dependence of the salmon catch on the two first principal components was studied with the regression analysis.

Results

The salmon catches have been large in low water temperatures in October (Table 1). There is a positive correlation between the salmon offshore catch in the Bothnian Sea and the water temperature in July of the previous year (Table 1).

The null hypothesis, that the data are samples from a normal distribution, cannot be rejected (Table 2).

The absolute correlations between the salmon off-

Table 1. Correlation of offshore catches of salmon in the Bothnian Sea with monthly mean water temperature 1974–1989.

	Actual year		Previous year	
Depth (m)	July	October	July	October
0	–0.19	–0.66	0.09	0.11
5	–0.16	–0.69	0.31	0.07
10	–0.12	–0.66	0.42	0.06
15	–0.26	–0.69	0.32	0.03
20	–0.40	–0.67	0.07	0.06

shore catch in the Bothnian Sea and the regressor variables are highest with the herring biomass, mean water temperature in October, number of real ice days, and number of group 1 herring (Table 3). The correlation with the actual herring year class is small and the variable was excluded from further analyses.

According to the coefficients of determination of the linear regression models with varying independent variables, the proportion of the salmon catch variation explained by the herring biomass and the smolt production of the previous year is 89% (Table 4). If the number of independent variables in the model is increased from 2 to 6, the coefficient of determination increases by 3%. Thus, in the material analysed, the herring biomass and smolt production of the previous year seem to be sufficient to explain the greater part of the yearly variation of the salmon catch in offshore fishing in the Bothnian Sea (Fig. 5).

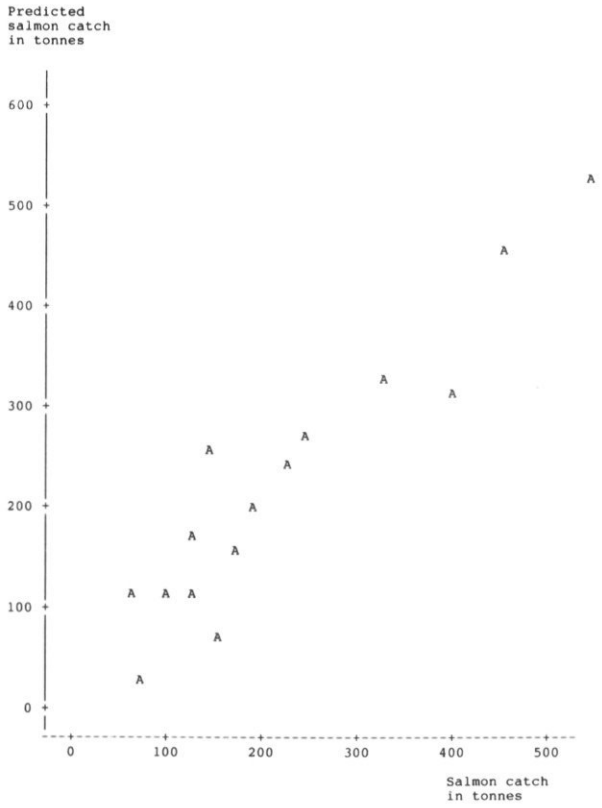

Figure 5. Predicted salmon catch in the offshore fishery in the Bothnian Sea (ICES Subdivision 30) according to the linear regression model based on herring biomass and smolt production.

Table 2. Means and standard deviations of the salmon, herring, and hydrographic variables in 1976–1990 and the probability that the characters are samples from a normal distribution ($n = 15$).

	Mean	s.d.	Normal probability
Salmon catch (t)	224.3	145.3	0.053
Smolt production (thousands)	3022	859	0.375
Herring year class	3972	2543	0.143
Group 1 herring (millions)	4014	2572	0.153
Herring biomass (1000 t)	423.9	106.1	0.061
Number of ice days	103.7	42.6	0.428
October 5 m, °C	7.6	1.1	0.548
Previous July 10 m, °C	9.76	1.9	0.090

The equation of the regression model is:

$$\text{Salmon catch (t)} = 62.07 + 1.59 * \text{herring biomass (t)} - 0.17 * \text{smolt production (in thousands)}.$$

The model suggests that a large herring stock biomass

Table 3. Correlations between salmon, Baltic herring abundance, and hydrographic factors in 1976–1990 ($n = 15$).

	Salmon catch	Smolt production	Herring year class	Group 1 herring	Herring biomass	Number of ice days	October 5 m, °C	Previous July 10 m, °C
Salmon catch	1.000							
Smolt production	−0.270	1.000						
Herring year class	−0.084	0.580	1.000					
1-group herring	0.425	0.351	0.211	1.000				
Herring biomass	0.530	0.633	0.288	0.603	1.000			
Number of ice days	−0.448	−0.350	−0.480	−0.612	−0.629	1.000		
October 5 m, °C	−0.499	0.490	0.217	−0.061	−0.021	−0.040	1.000	
Previous July 10 m, °C	0.418	−0.087	−0.058	0.166	0.157	−0.263	0.014	1.000

Table 4. The biggest coefficients of determination (R^2) and the error sum of squares of the regression models between the salmon offshore catch in the Bothnian Sea and various regressor variables in 1976–1990.

No. of regressor variables in the model	Variables	R^2	Error sum of squares
1	Herring biomass	0.281	212 603
2	Herring biomass Smolt production	0.889	32 657
3	Herring biomass Smolt production Previous July 10 m, °C	0.913	25 729
4	Herring biomass Smolt production Previous July 10 m, °C Group 1 herring Number of ice days	0.919	23 795
5	Herring biomass Smolt production Previous July 10 m, °C Group 1 herring Number of ice days	0.920	23 564
6	Herring biomass Smolt production Previous July 10 m, °C Group 1 herring Number of ice days October 5 m, °C	0.920	23 516

in the actual year and small smolt production in the previous year predict large offshore salmon catches. Under such circumstances a maximum amount of herring is available per salmon, and a large proportion of salmon may stay in the Bothnian Sea, instead of extending the feeding migration to the Baltic Main Basin.

Principal components were computed from the smolt production, herring abundance, and hydrographical variables (Table 5). The first component accounts for 45% of the variance and the first two components explain 67%. According to the eigenvectors (Table 5), the second principal component is mainly a measure of sea temperature and smolt production and the first component a measure of other characters.

Table 5. The eigenvectors and the cumulative variance of the first three principal components based on smolt production, herring abundance, and hydrographic factors in the Bothnian Sea in 1976–1990.

	Principal components		
	1	2	3
Smolt production	0.435	0.514	−0.087
Group 1 herring	0.482	−0.237	−0.159
Herring biomass	0.539	−0.056	−0.181
Number of ice days	−0.500	0.204	−0.052
October 5 m, °C	0.117	0.701	0.465
Previous July 10 m, °C	0.155	−0.377	0.849
Eigenvalue	2.672	1.362	0.968
Cumulative variance	0.445	0.672	0.834

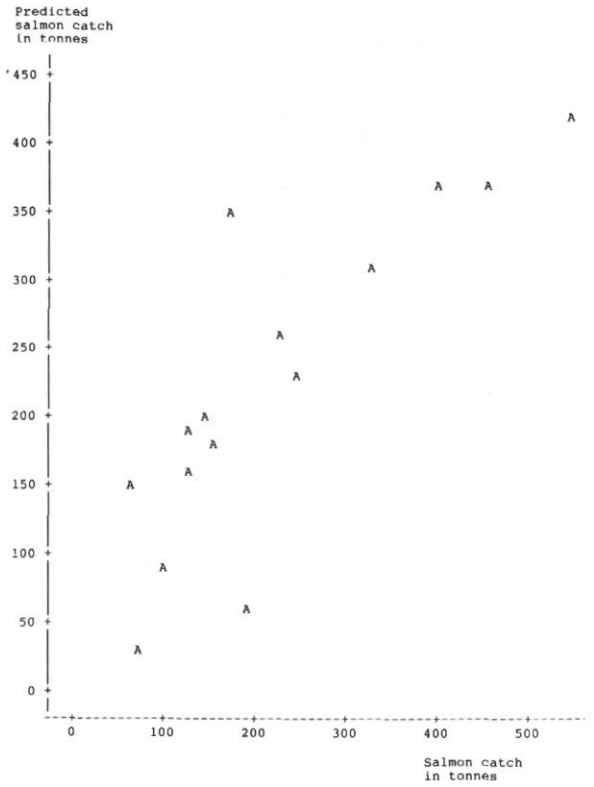

Figure 6. Predicted salmon catch in the offshore fishery in the Bothnian Sea (ICES Subdivision 30) according to the linear regression model based on the first two principal components of the smolt production, herring abundance, and hydrographical variables.

According to the coefficients of determination of the linear regression model, the first two principal components explain 69% of the yearly salmon catch variation in offshore fishing in the Bothnian Sea (Fig. 6).

Discussion

The regressor variables used in the models obviously do not predict the salmon catches, but the amount of salmon in the Bothnian Sea. The salmon catch is used in the model because yearly data on the amount of salmon in the sea are lacking. C.p.u.e. data are available for the salmon drift net and long line records made in the Bothnian Sea in the period 1980–1989. Their correlations (r) with the offshore salmon catch in the Bothnian sea are 0.74 and 0.53, respectively.

The hypothesis that the offshore salmon catch in the Bothnian Sea is connected with the feeding conditions is

supported by a positive correlation (r = 0.67) between the catch and the mean weight of salmon age group A.1+ in the Bothnian Sea in October–December 1976–1990.

In 1991 the herring stock biomass in the Gulf of Bothnia is 720 000 tonnes (Anon., 1991b) and the smolt production of the year 1990 was 3 290 000 (Fig. 2). According to the first model presented, a salmon catch of 650 tonnes can be predicted in the offshore fishery in the Bothnian Sea in 1991. The predicted catch is higher than earlier, which is in accordance with the data obtained from the fishery, where catches have been unusually high in the Bothnian Sea in January–May 1991.

As presented earlier the abundance of the herring population and number of group 1 herring seem to be the main variables required to predict the salmon catches in the Bothnian Sea. During the post-smolt run, this herring year class is at the age of 0+. As herring spawn in coastal waters, the migrating post-smolts and 0+ herring occur in the same area.

A possible indication of a large number of salmon feeding in the Bothnian Sea may be an abundance herring year class. In these circumstances a larger proportion of salmon post-smolts than average prefer to stay in the Bothnian Sea, utilizing herring which were born in the same spring when the smolts entered the sea, and whose size therefore remains suitable as the food of the growing post-smolts.

If the herring biomass is small, the feeding conditions are poor and the post-smolts prefer to migrate to the Baltic Main Basin, where food is always available. The correlation between low smolt production and good landings is difficult to explain. A possible reason may be competition for food. However, as the annual changes in the number of post-smolts are quite small compared with those in the herring biomass, the post-smolts must have a negligible effect on the herring biomass.

More data are needed in order to clarify the interactions between salmon, herring, and hydrographical factors.

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ICES mar. Sci. Symp., 195: 499–506. 1992

Variability in the nutrient reserves and pelagic productivity in the western Gulf of Finland

Lars Grönlund and Juha-Markku Leppänen

Grönlund, L., and Leppänen, J.-M. 1992. Variability in the nutrient reserves and pelagic productivity in the western Gulf of Finland. – ICES mar. Sci. Symp., 195: 499–506.

Long-term data on hydrography, major nutrients, chlorophyll *a* and phytoplankton primary production have been collected frequently since 1972 at a sampling station off the Tvärminne archipelago at the entrance to the Gulf of Finland. The data are evaluated in order to determine long-term trends and variability in the nutrient reserves and plankton productivity during the different growth periods. A significant shift in the frequency distribution of the nutrient concentrations towards higher values was observed in the 1980s. Productivity and variability in the pelagic system have increased predominantly during the vernal growth period.

Lars Grönlund and Juha-Markku Leppänen: Finnish Institute of Marine Research, PO Box 33, SF-00931 Helsinki, Finland.

Introduction

The aim of this paper is to evaluate the long-term data collected by the Finnish Institute of Marine Research at a coastal site in order to determine whether changes in nutrient concentrations and phytoplankton productivity and biomass have occurred during the period 1972–1988.

The sampling site, which is situated in the open sea area at the entrance to the Gulf of Finland, is affected by both local and long-distance inputs of nutrients. Nutrient-rich deep water is able to flow from the Baltic Proper into the Gulf of Finland, and upwellings are common near the sampling area. The Gulf of Finland is affected by both land-based and atmospheric pollution (HELCOM, 1987). Most of the land-based load is discharged into the eastern part of the Gulf of Finland (Pitkänen *et al.*, 1987). Owing to the coriolis effect, this surface water sweeps along the Finnish coast. The study area is also characterized by an occasional outflow of almost fresh water from the adjacent Pojo Viken area.

Material and methods

The sampling station is situated in the outer archipelago near the entrance to the Gulf of Finland (Fig. 1). Hydrography, nutrients, chlorophyll *a* and phytoplankton primary production have been measured almost biweekly at this station since 1972. The location of the sampling station was changed in 1976 to a point ca. 5 km closer to the coast (Fig. 1, Stations 1a and 1b). According to Laakkonen *et al.* (1981), both sampling stations represent the same watermasses.

The total number of observations during the period 1972–1988 was ca. 290. Winter sampling has been sparse and irregular, most of the sampling being done during the ice-free period. The following data analysis is therefore concentrated on the phytoplankton growth season, between March and August. The main attention is paid to total phosphorus and nitrogen, chlorophyll *a* and daily primary production.

The analysis methods of Koroleff (1979) have been used for total phosphorus and nitrogen.

The samples for daily primary production were incubated *in situ* for 24 h as described by Lassig and Niemi (1972). Chlorophyll *a* was determined spectrophotometrically up until 1978 (Niemi *et al.*, 1970), and since then fluorometrically, as described in Edler (1979). The slight modifications of the methods have not affected the results.

Depth-balanced, integrated average values from 0 to 10 m were used in the data processing. The data were also divided into two different growth seasons. A spring bloom season comprising data from March, April, and May, and a summer period comprising June, July, and August.

In order to illustrate changes that have occurred during the two consecutive decades, the data were

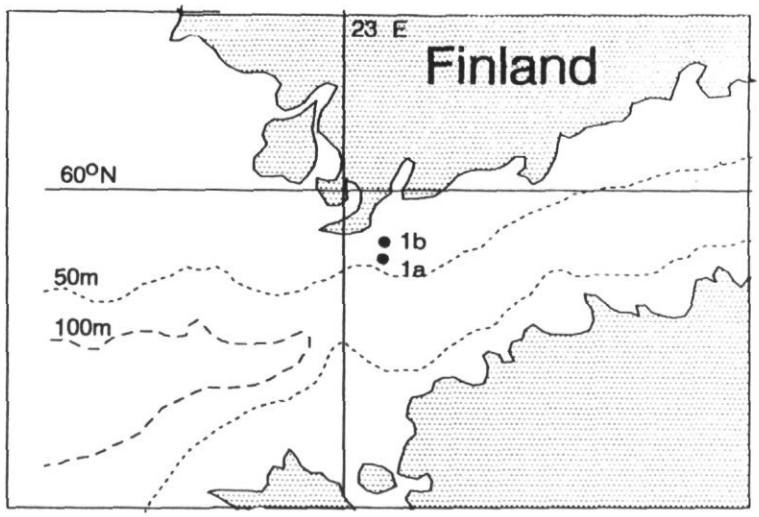

Figure 1. Location of the sampling station. The station was moved to a point ca. 5 km nearer the coast in 1976 (marked as 1b).

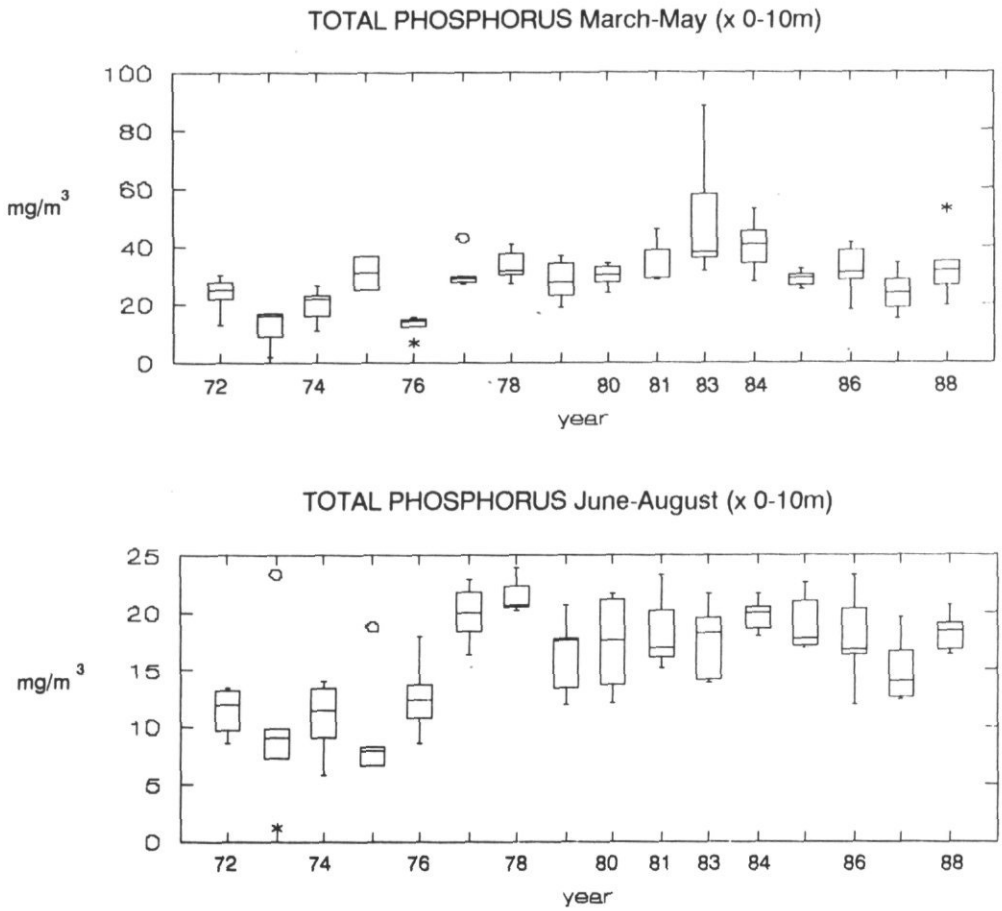

Figure 2. The interannual variability of total phosphorus concentrations in the 0–10 m water layer off Tvärminne archipelago during spring and summer. The central box covers the middle 50% of the data values, between the upper and lower quartiles of the respective year. The "whiskers" extend to the minimum and maximum values, and the central line is at the median of the annual values. Extreme values are plotted as separate points.

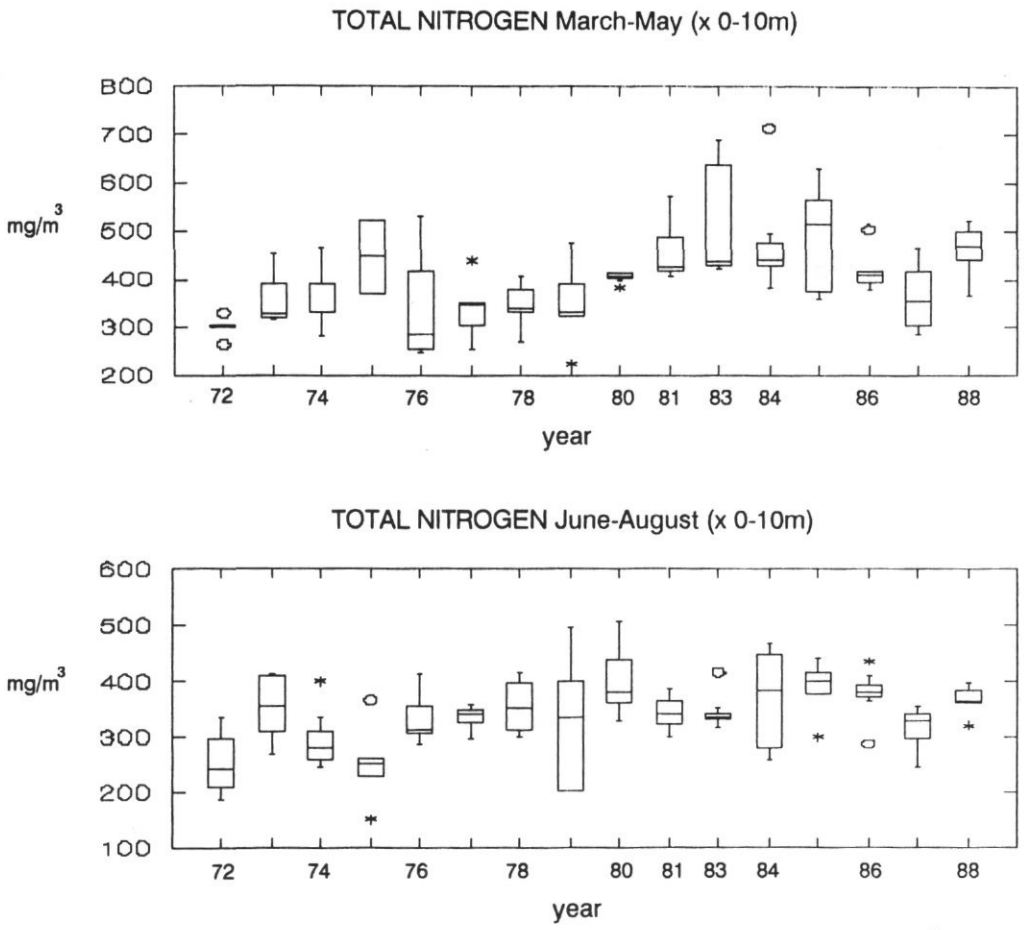

Figure 3. The interannual variability of total nitrogen concentrations in the 0–10 m water layer off Tvärminne archipelago during spring and summer. The central box covers the middle 50% of the data values, between the upper and lower quartiles of the respective year. The "whiskers" extend to the minimum and maximum values, and the central line is at the median of the annual values. Extreme values are plotted as separate points.

divided into a period stretching from 1972 to 1979 and another from 1980 to 1988.

A Kolmogorov–Smirnov two-sample test for non-linear samples was used to test the differences between the decades.

Results

The total phosphorus concentrations increased in the late 1970s during both the vernal and summer period (Fig. 2). However, there was a similar rise in the total nitrogen concentration only in the spring period during the 1980s (Fig. 3). For chlorophyll *a*, both the levels and variation in concentrations have grown during spring and summer from 1983 onwards (Fig. 4). The results for daily primary production did not show an increase either in level or in variation (Fig. 5). The rise in the chlorophyll *a* concentrations coincides with a decrease in salinity from 1983 onwards (Fig. 6). This drop in the salinity level is not as clear in the summer values as it is during spring.

The frequency distributions of different concentration classes for vernal and summer total phosphorus during the two periods 1972–1979 and 1980–1988 are shown in Figure 7. They indicate a significant shift towards larger values during the 1980s (Table 1). A similar shift in the values for total nitrogen was seen in the spring values especially (Fig. 8 and Table 1). A change in the chlorophyll *a* values was evident during the spring especially (Fig. 9 and Table 1). The Kolmogorov–Smirnov test on the daily primary production values showed no significant differences for the summer values, but the difference for the spring was significant at about the 1% risk level (Table 2 and Fig. 10).

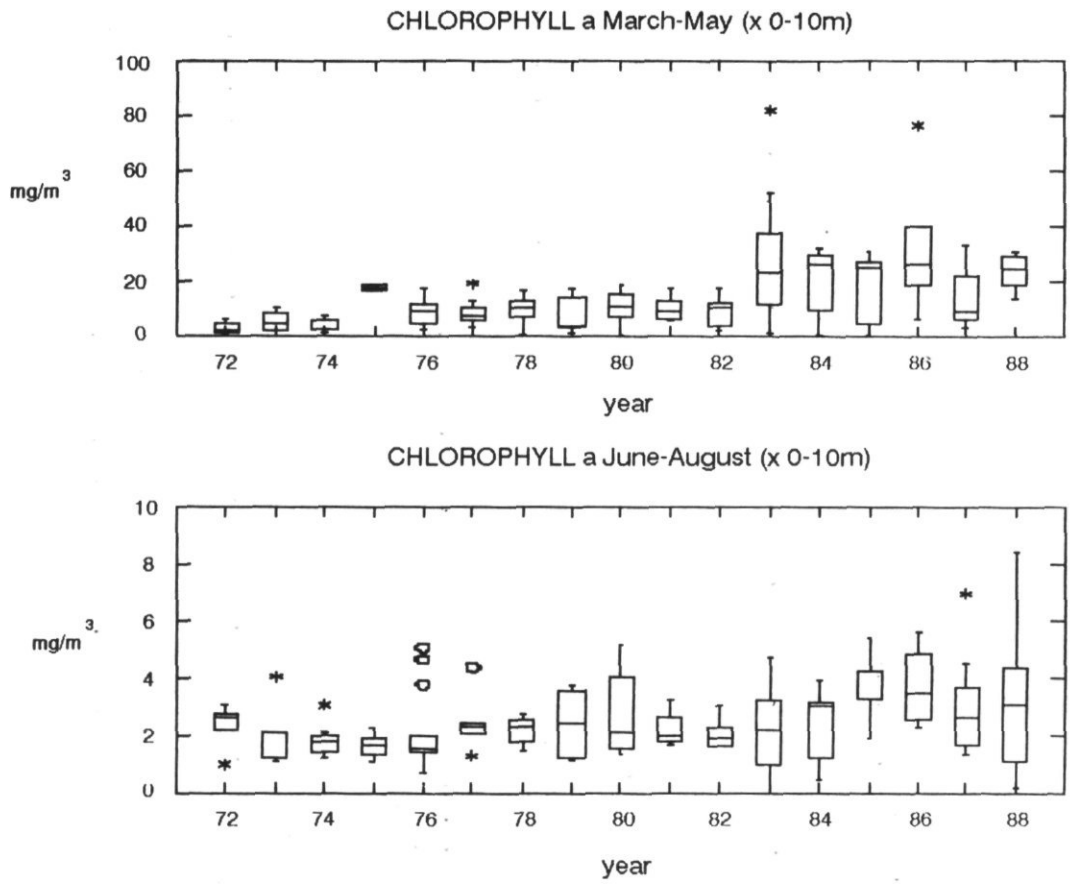

Figure 4. The interannual variability of chlorophyll *a* concentrations in the 0–10 m water layer off Tvärminne archipelago during spring and summer. The central box covers the middle 50% of the data values, between the upper and lower quartiles of the respective year. The "whiskers" extend to the minimum and maximum values, and the central line is at the median of the annual values. Extreme values are plotted as separate points.

Table 1. Some statistical characteristics for total nitrogen (totN), total phosphorus (totP), chlorophyll *a* (Chl *a*), and daily primary production (PP) for the growing periods 1972–1979 and 1980–1988 respectively. *n* refers to sample size and *x* to average, median, maximum, minimum and standard deviation (s.d.) are also given. The units are mg m^{-3} except for daily primary production which is given as mgC m^{-2} d^{-1}.

	Period	*n*	*x*	Median	Max	Min	s.d.
SPRING							
TotN	72–79	41	348	331	530	222	72.7
	80–88	49	447	422	715	287	94.6
TotP	72–79	40	24.3	25.6	43.1	2.2	9.6
	80–88	49	34.1	31.8	88.6	15.0	12.8
Chl *a*	72–79	43	7.9	7.0	19.3	0.7	5.5
	80–88	62	19.6	14.9	82.6	0.5	16.6
Prim.	72–79	33	701	699	2023	2.3	361
prod.	80–88	42	899	785	2221	116	531
SUMMER							
TotN	72–79	48	315	312	499	151	70.0
	80–88	49	363	363	510	245	54.1
TotP	72–79	48	13.4	13.0	24.0	1.2	5.4
	80–88	50	17.7	18.2	23.3	12.0	3.1
Chl *a*	72–79	52	2.1	1.9	5.1	0.7	1.0
	80–88	62	2.8	2.7	8.4	0.1	1.7
Prim.	72–79	51	419	398	1053	146	194
prod.	80–88	50	335	311	787	33.9	148

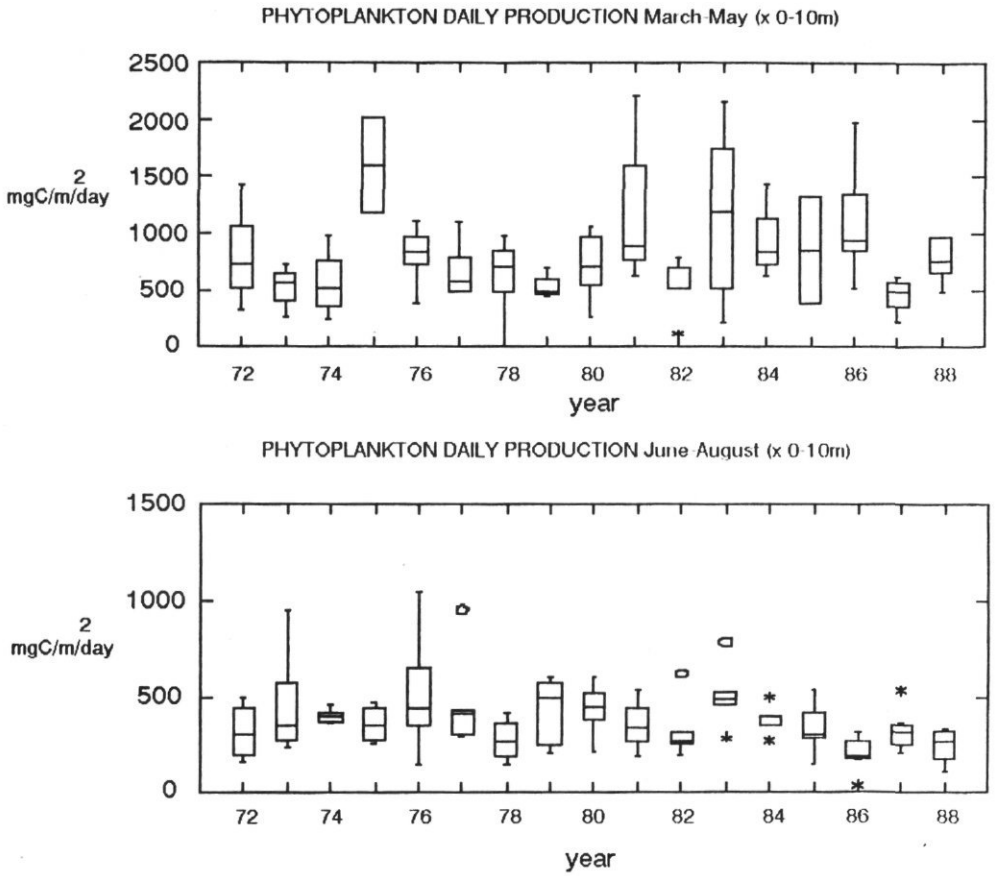

Figure 5. The interannual variability of phytoplankton daily primary production values in the 0–10 m water layer off Tvärminne archipelago during spring and summer. The central box covers the middle 50% of the data values, between the upper and lower quartiles of the respective year. The "whiskers" extend to the minimum and maximum values, and the central line is at the median of the annual values. Extreme values are plotted as separate points.

Discussion

Several reports about increasing nutrient concentrations, changes in the composition of the phytoplankton communities and exceptional blooms that are occasionally toxic have been reported from the coastal areas around the Gulf of Finland (Kangas *et al.*, 1982; Kononen and Niemi, 1984, 1986; Persson *et al.*, 1984; Ilus and Keskitalo, 1987; Pitkänen *et al.*, 1987; Järvekülg *et al.*, 1988; Kononen, 1988; Sivonen *et al.*, 1989; HELCOM, 1990).

The extremely high chlorophyll *a* spring peaks in 1983 and 1986 were related to the freshwater outflow from Pojo Viken, creating stable conditions for the blooms by stabilizing the upper layer and adding an excess of nutrients. The decrease or depletion of the reserves of reactive silicate in the euphotic layer after the spring bloom (Niemi and Åström, 1987), which is a relatively recent observation, indicates enhanced diatom production at that time.

The daily production values fluctuated, probably in accordance with the light conditions prevailing during the *in situ* incubation, and changes in the variability of the values for the different decades could only be detected during the spring bloom.

The increase in the total phosphorus concentration during the summer period is probably caused by the increase in organic phosphorus, since the levels of inorganic phosphorus have not increased (unpublished data

Table 2. Levels of significance for the different parameters during spring and summer.

	Spring	Summer
TotN	<0.001	0.0038
TotP	<0.001	0.0014
Chl *a*	<0.001	0.0067
Prim. prod.	0.017	0.4020

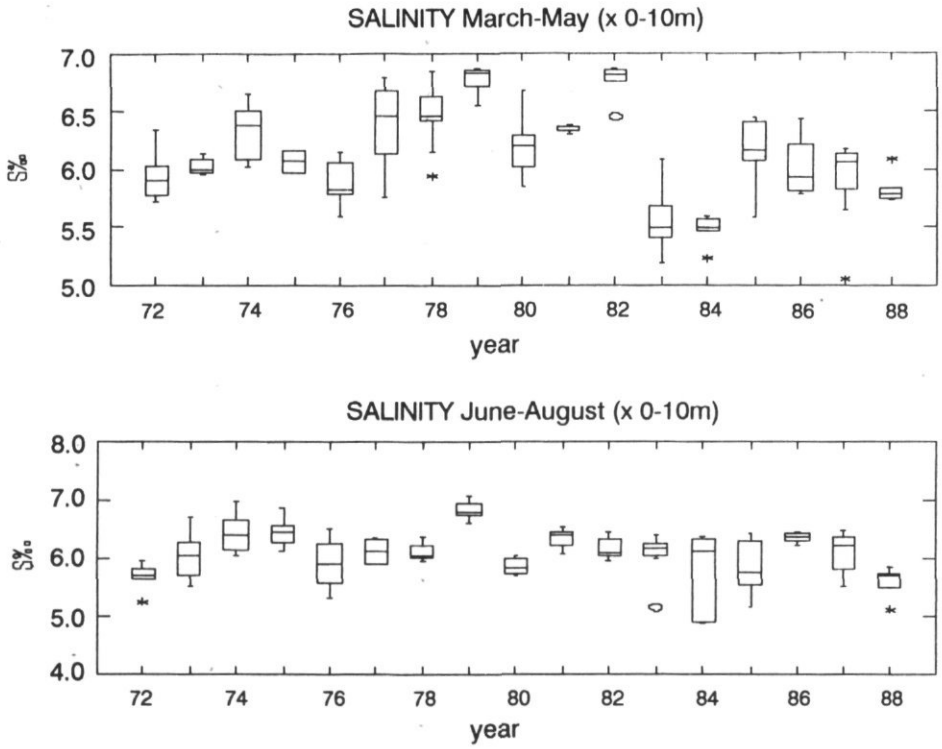

Figure 6. The interannual variability of salinity concentrations in the 0–10 m water layer off Tvärminne archipelago during spring and summer. The central box covers the middle 50% of the data values, between the upper and lower quartiles of the respective year. The "whiskers" extend to the minimum and maximum values, and the central line is at the median of the annual values. Extreme values are plotted as separate points.

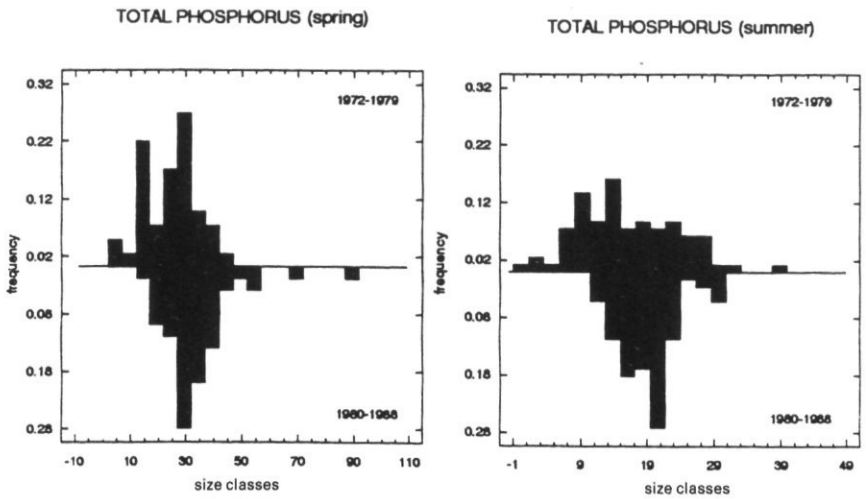

Figure 7. Frequency histogram for the total phosphorus concentrations during the spring and summer months respectively. The bars above the horizontal line represent the first eight years of the monitoring period, and those below the last nine years of the period 1972–1988.

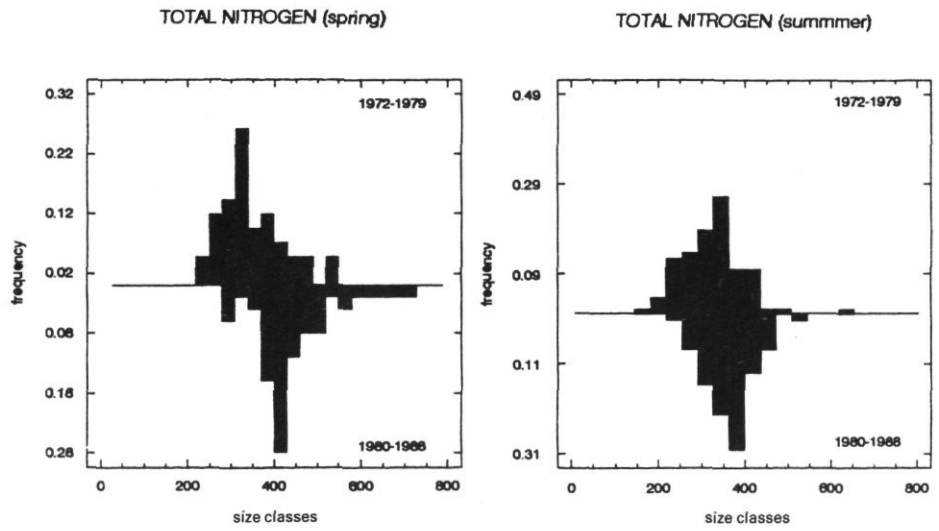

Figure 8. Frequency histogram for the total nitrogen concentrations during the spring and summer months, respectively. The bars above the horizontal line represent the first eight years of the monitoring period, and those below the last nine years of the period 1972–1988.

of the Institute of Marine Research). Kononen (1988) found that some nanoplanktonic species, as well as a heterotrophic species, increased in abundance at the same area during the summer months in 1972–1985. This coincided with changes in the values for total phosphorus.

The present data show that the pelagic system in the outer archipelago of the western Gulf of Finland has reached a stage where the increase in nutrient reserves is intensifying the algal blooms in spring. Extensive sedimentation of the blooms in spring, zooplankton grazing and rapid sedimentation of fecal pellets in summer, formation of refractory organic compounds, denitrification, and the binding of phosphorus in the sediments can, however, still counteract the eutrophication process. The present data also indicate that the increasing tendency in the nutrient and chlorophyll *a* concentrations has not continued in the 1980s.

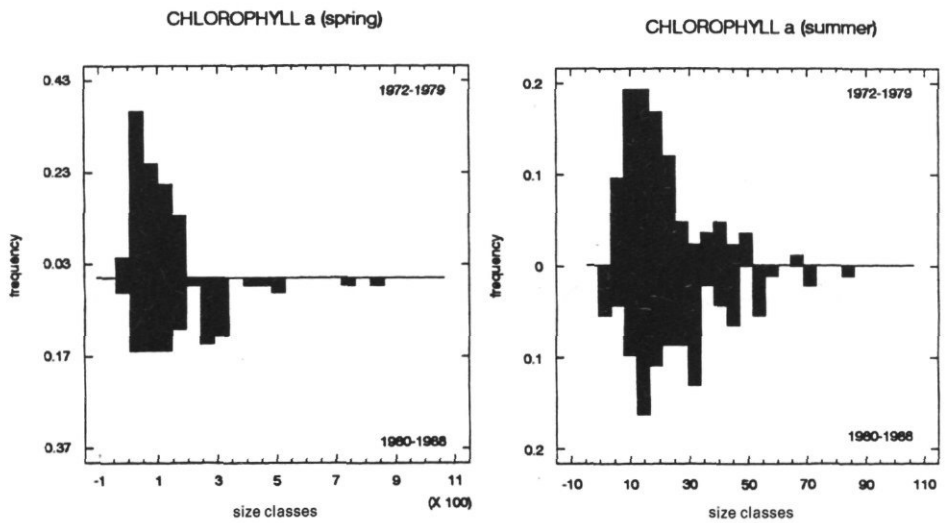

Figure 9. Frequency histogram for the chlorophyll *a* concentrations during the spring and summer months respectively. The bars above the horizontal line represent the first eight years of the monitoring period, and those below the last nine years of the period 1972–1988.

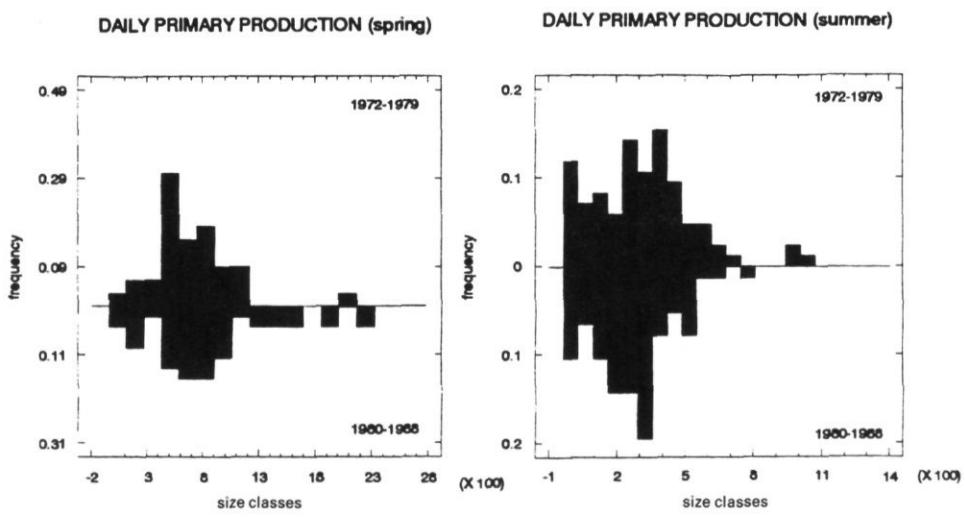

Figure 10. Frequency histogram for the phytoplankton daily primary production values during the spring and summer months. The bars above the horizontal line represent the first eight years of the monitoring period, and those below the last nine years of the period 1972–1988.

Acknowledgments

The authors thank Julius Lassig, Fil.lis., and Professor Åke Niemi for their farsightedness in starting these studies. We also thank the National Board of Waters and Environment for financing the coastal study during the last few years. Valuable help has been given by the personnel of the Tvärminne Zoological Station.

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ICES mar. Sci. Symp., 195: 507–514. 1992

Inorganic phosphorus and nitrogen compounds as driving forces for eutrophication in semi-enclosed seas

Dietwart Nehring

Nehring, D. 1992. Inorganic phosphorus and nitrogen compounds as driving forces for eutrophication in semi-enclosed seas. – ICES mar. Sci. Symp., 195: 507–514.

Eutrophication is identified as one of the most serious problems in the world. Trend studies covering the Baltic Proper in the period 1958–1989 show on average increasing phosphate and nitrate concentrations in the surface layer in winter and in the deep water below the permanent halocline. The accumulation rates have, however, been decreasing recently, except for phosphate in the anoxic deep water. Consequences of eutrophication are discussed with respect to biological productivity and deterioration of the oxygen conditions in Baltic deep waters. Conditions for eutrophication in the Baltic Sea are compared with those in the Mediterranean Sea.

Dietwart Nehring: Baltic Sea Research Institute, D-2530 Warnemünde, Germany.

Introduction

Eutrophication has developed into a global problem which increasingly affects coastal seas and semi-enclosed basins. This process is mainly caused by land-based and airborne inputs of nutrients of anthropogenic origin (Larsson *et al.*, 1985), whereas natural sources are of low significance in this connection. Eutrophication also depends on the water exchange with the adjacent seas.

Results

Hydrographic monitoring programmes have been carried out in the Baltic Sea since the end of the nineteenth century. Nutrients and biological parameters have been included in these programmes in recent decades. Figure 1 is a map showing the monitoring stations in the Baltic Proper used in the following discussion.

The nutrient data continuously measured since the late 1950s enable long-term studies on the development of the nutrient situation in the Baltic Proper (Nehring and Matthäus, 1991). The results explained for stations in the eastern Gotland Basin are also valid for other subregions of the Baltic Sea Area (Baltic Marine Environment Protection Commission, 1987).

Figure 2 shows that the phosphate and nitrate concentrations are increasing, on average, in the surface layer in winter, when the light limits phytoplankton development. The sub-trends indicate that the positive overall trends mainly result from the strong increase in the period 1969–1977, whereas nutrient concentrations are characterized by pronounced fluctuations in recent times, but no longer increase, on average. Some similarities exist between the trend behaviour of the nutrients and the salinity (Fig. 2), indicating that eutrophication in the Baltic ecosystem also benefits from the advective water renewal, transporting deep water rich in nutrients and salinity into the surface layer.

Increasing nutrient concentrations have also been observed in the oxic deep waters below the halocline, as shown in Figure 3. The split nutrient trends do not differ very much from the positive overall trends. The mean annual accumulation rate of phosphate, however, is characterized by a significant decrease in the second period, whereas nitrate accumulation continues virtually unchanged throughout the period under investigation. Oxygen conditions have the tendency to improve in this layer, although no significant trend could be identified. The trend of salinity is significantly positive at the beginning of the investigations and negative since the late 1970s. Because of this inverse behaviour, the overall trend strongly differs from the sub-trends.

The situation under alternating oxic and anoxic conditions is illustrated for the 200 m depth in Figure 4. Nitrate was excluded from the investigations because of denitrification. Hydrogen sulphide was converted into negative oxygen concentrations ($S^{2-} + 2O_2 = SO_4^{2-}$) in order to get comparable parameters.

The most important feature of the parameters studied in this depth is the different trend behaviour in the separate periods under investigation. In the beginning,

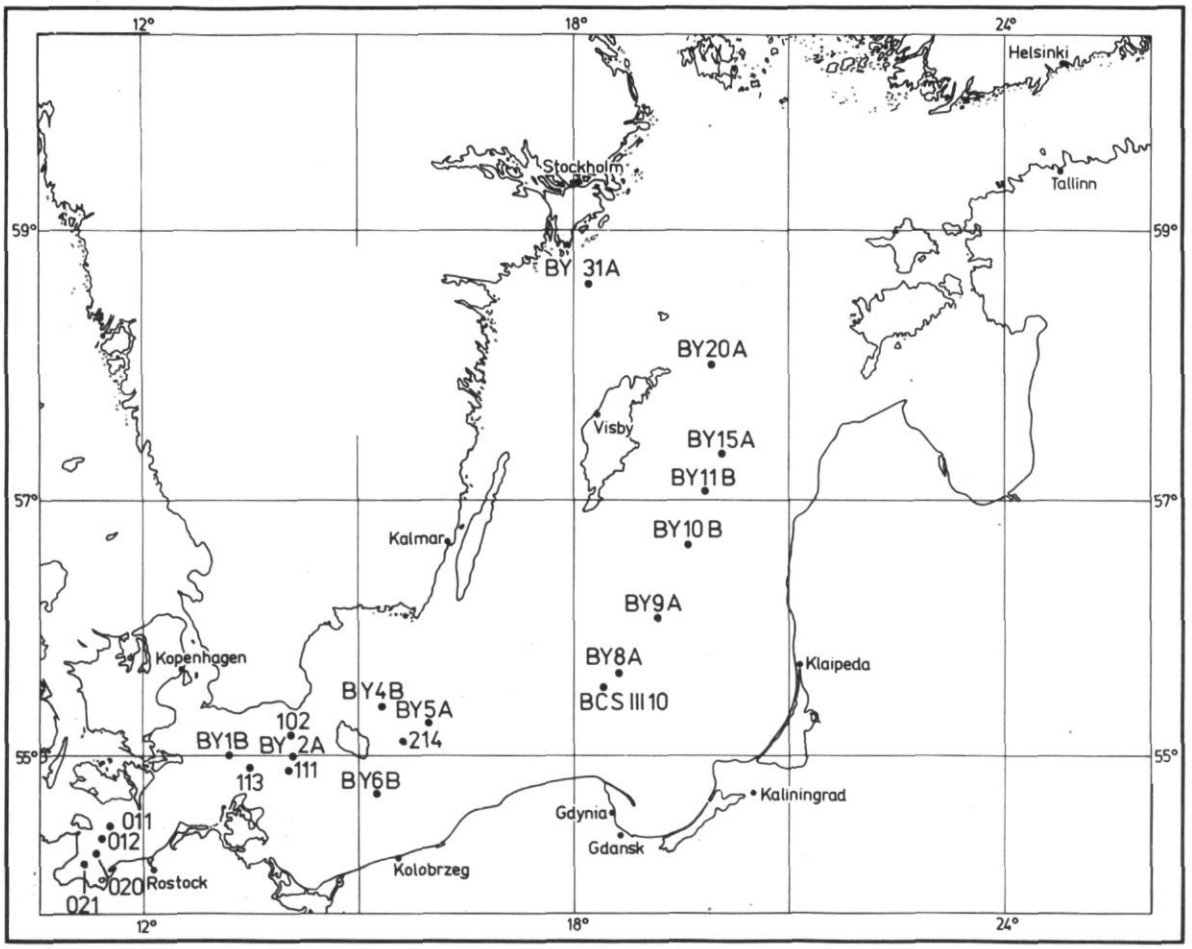

Figure 1. Map of stations.

phosphate, oxygen, and salinity are characterized by moderate negative trends. The concentrations of the parameters strongly fluctuate in this period in the sequence of water renewals and stagnant conditions.

Since 1977 stagnant conditions have prevailed, causing a strong and almost continuous decrease in oxygen and salt concentrations, whereas phosphate concentrations have increased mainly due to phosphate remobilization from the sediments in the presence of hydrogen sulphide (cf. Nehring, 1987).

In comparison with the eastern Gotland Basin, the deep water is more often renewed in the western basins of the Baltic Proper, i.e. the Bornholm Basin and the Gdańsk Basin. The deep water of the Arkona Basin is annually exchanged in late autumn. Figure 5 shows the distribution of hydrochemical parameters in the near-bottom water layer of the Bornholm Deep. Variations caused by water renewals and stagnation periods superimpose the long-term changes which are, however, significant.

Discussion

Phosphate and nitrate trends were updated for the surface layer and the deep water in the Baltic Proper (cf. Nehring *et al.*, 1989). Winter concentrations of these nutrients are no longer increasing, on average, in the surface layer, whereas the overall trends covering the periods 1958 to 1989 for phosphate and 1969 to 1989 for nitrate are still positive. Since there is no indication of considerable changes in potential land-based and airborne inputs into the Baltic Sea the following may be occurring: (1) A near steady state may have developed between nutrient sources and sinks. (2) Interannual 3 to 4 and 6 to 7 year cycles which are attributed to changes in the atmospheric circulation create variations in the river discharge and are also reflected by changes in the nutrient trends (Trzosińska, 1990).

Although there is no further increase, on average, the recent nutrient pool involved in the biological cycle is at such a high level that the biogenic material in the course

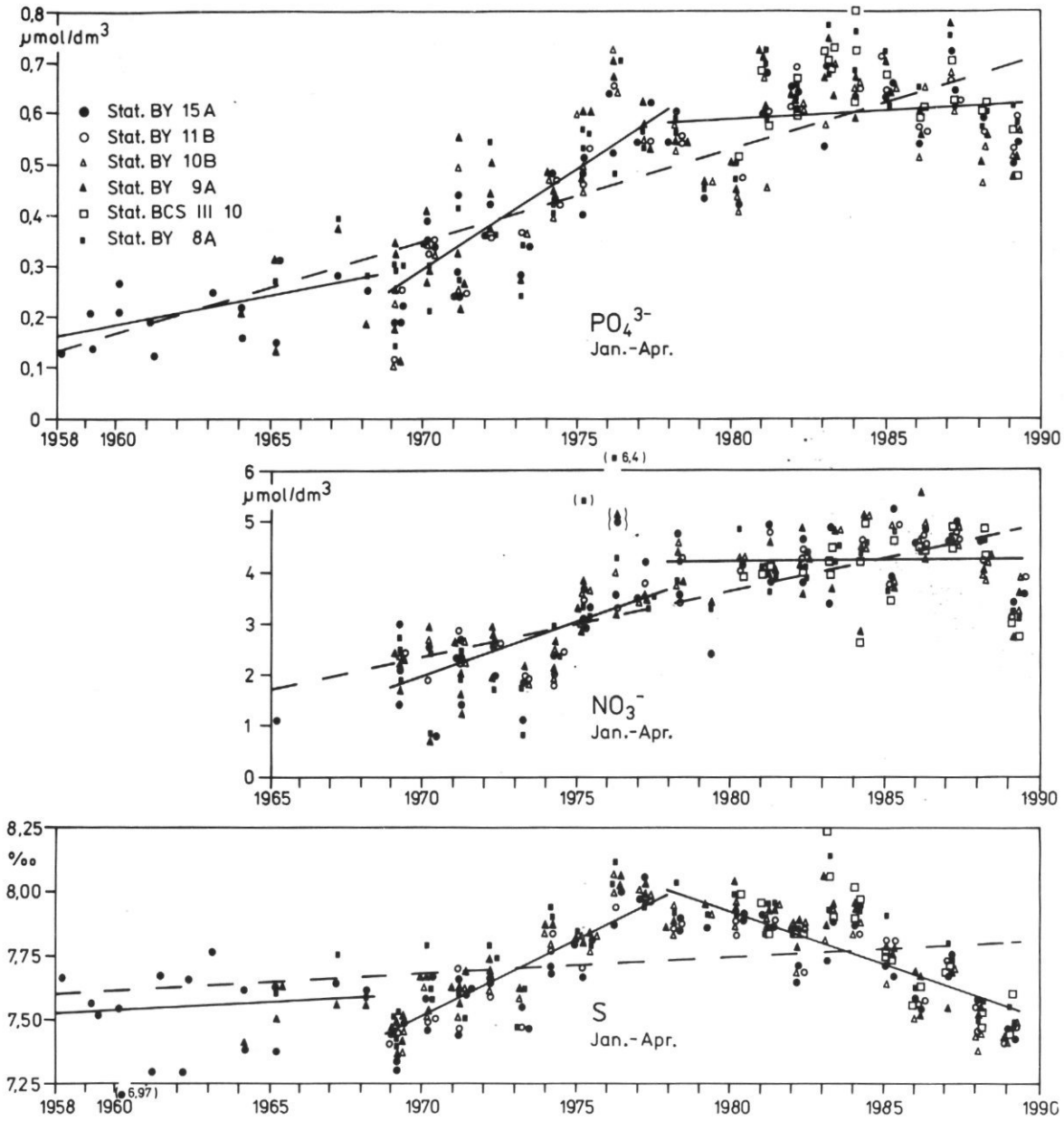

Figure 2. Long-term trends of nutrients and salinity in the surface layer in winter at stations in the Central Baltic Sea.

of its microbial destruction causes ongoing deterioration of the oxygen conditions in Baltic deep waters or at least in the transition zone between water and sediment.

Trend changes observed in the surface layer are also reflected to a certain extent in the oxic deep water of the Central Baltic basins. High phosphate accumulation rates have been identified in the near-bottom water layers of the Central Baltic deeps in connection with the recent stagnation period beginning in 1977. They mainly result from the remobilization of phosphate from the sediments, when the redox potential decreases by the development of anoxic conditions.

Absent major inflows of highly saline and oxygen-rich water are reflected by the decrease in oxygen and salt concentrations in Central Baltic deep waters. The recent stagnation period is the most serious that has ever been observed in the Gotland Deep up to now (Matthäus, 1990).

The continuous decrease in salinity is the reason that the near-bottom water layer in the Central Baltic basins

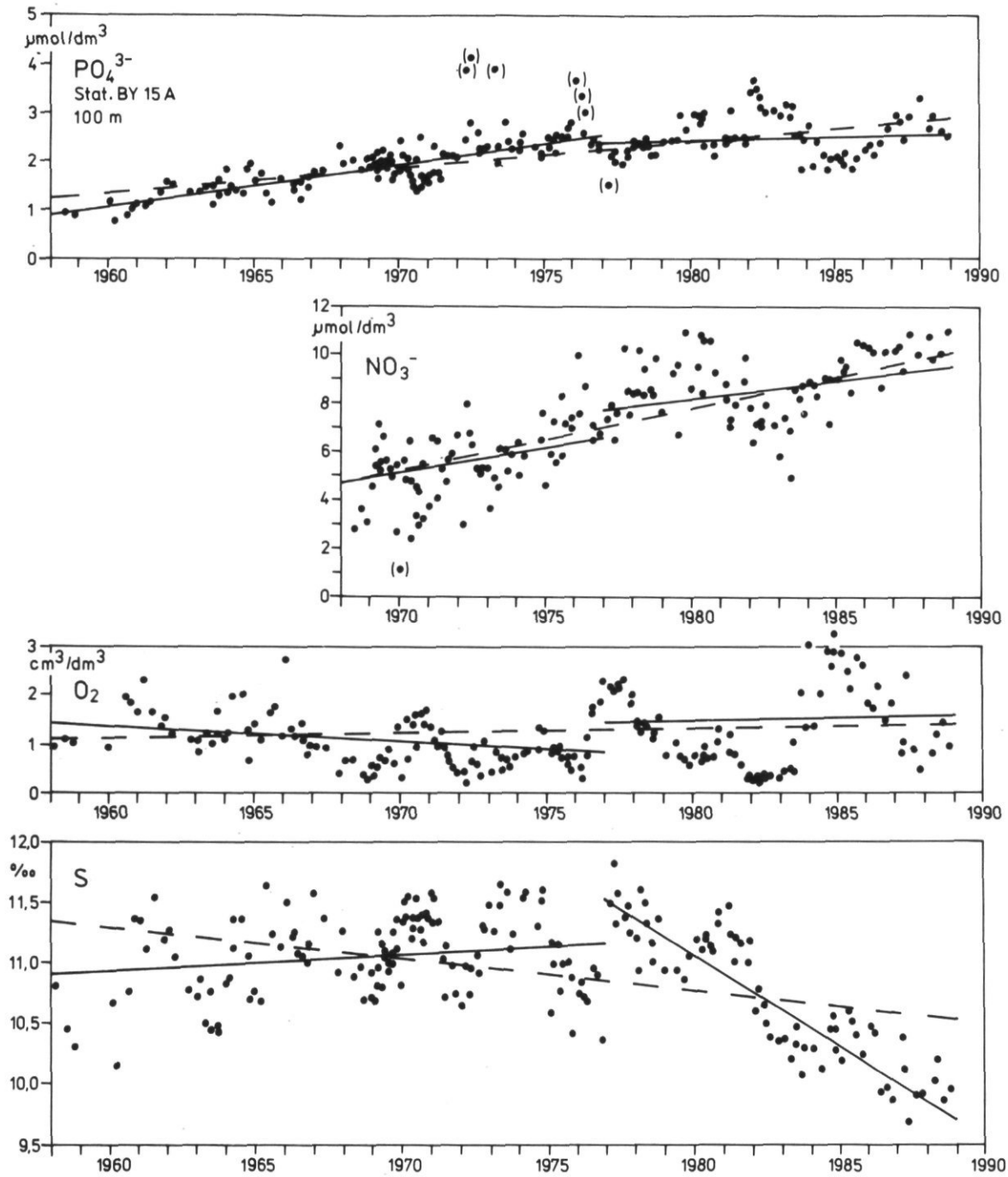

Figure 3. Long-term hydrographic and nutrient variations below the halocline at a Central Baltic station (Station 15A, 100 m depth). Oxic conditions.

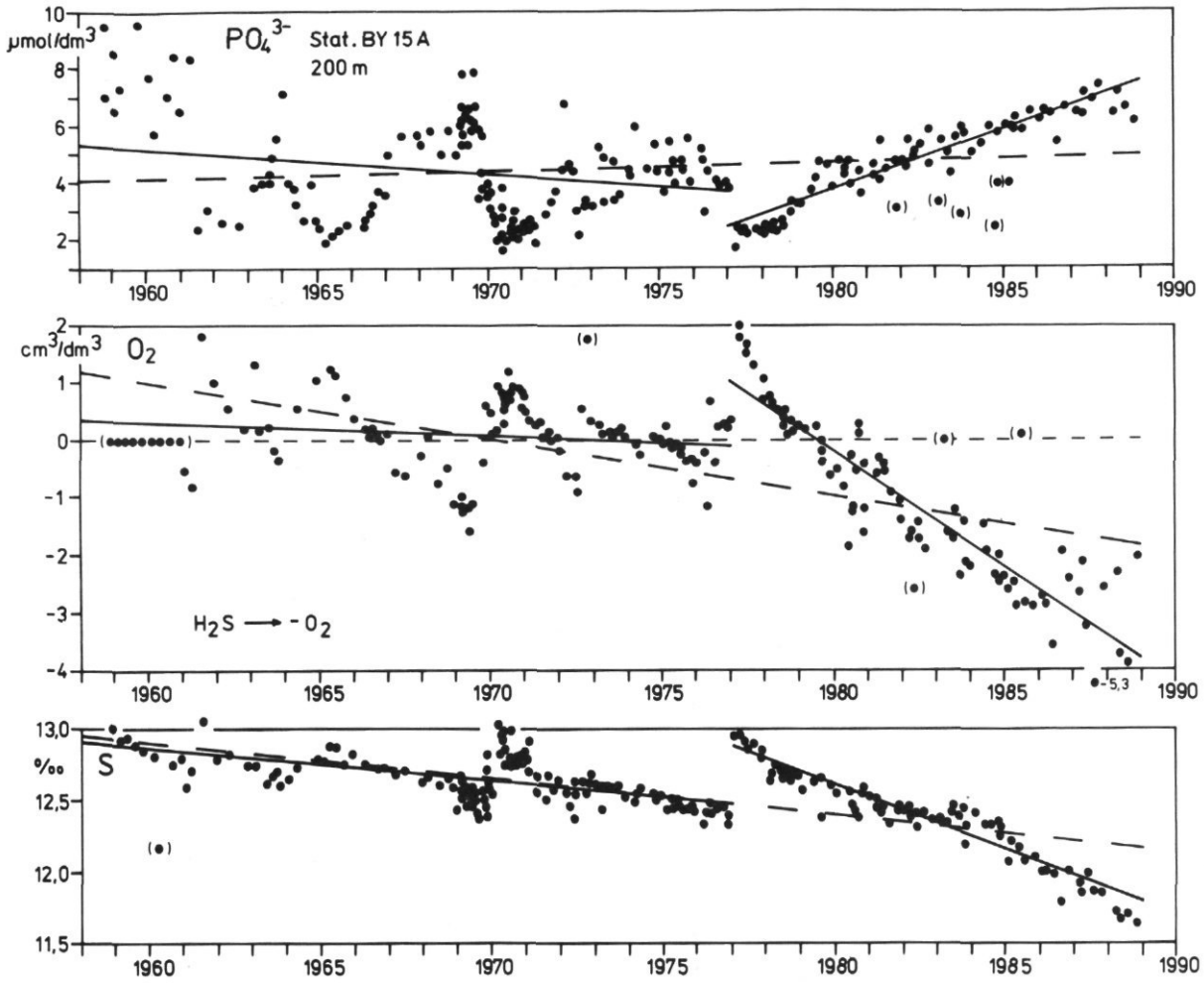

Figure 4. Long-term hydrographic and nutrient variations in the near-bottom water layer at a Central Baltic station (Station 15A, 200 m depth). Alternating oxic and anoxic conditions, hydrogen sulphide converted into negative oxygen equivalents.

is increasingly affected by advective processes and vertical mixing without major inflows. This is reflected by the more frequent renewal of the near-bottom water layer in the Bornholm Basin and in the Gdańsk Basin. The strong decrease in salinity observed in the latter basin in spring 1990 (Nehring, 1991) indicates, however, that the deep water can also be exchanged by wind-driven vertical mixing, facilitated by the recent weak haline stratification. The improvement in the oxygen conditions in the western Gotland Basin and variations in the hydrogen sulphide concentrations in the Gotland Deep (Nehring, 1991) are also consequences of advective processes favoured by the decreasing salinity in the deep water.

The inorganic phosphorus and nitrogen compounds are the driving forces in eutrophication. Using only winter concentrations, the overall phosphate and nitrate trends are significantly positive in the surface layer of the Baltic Sea indicating eutrophication. The high land-based and atmospheric inputs of nutrients from anthropogenic sources (Larsson *et al.*, 1985) show the influence of human activities on this process.

The most important consequence of the increasing or at least high nutrient pool in the Baltic Sea is the growing biological productivity. This is indicated by the positive trends in the phytoplankton biomass (Chl *a*) and primary productivity (Schulz and Kaiser, 1986), and also by the increasing biomass of the macro zoobenthos (Cederwall and Elmgren, 1980; Gosselck, 1985) as long as the oxygen concentration is not the limiting factor.

According to ICES statistics for the Baltic Sea, total catches of the most important commercial fish species herring, sprat, and cod have also increased from 450 000 t in 1966 to 900 000 t in 1980, remaining thereafter at this level. Not only the growing fishing effort but also eutrophication, bringing about better nutrition conditions for fish larvae and fish, is responsible for the increasing catches in the 1970s.

The most important negative change caused by eutro-

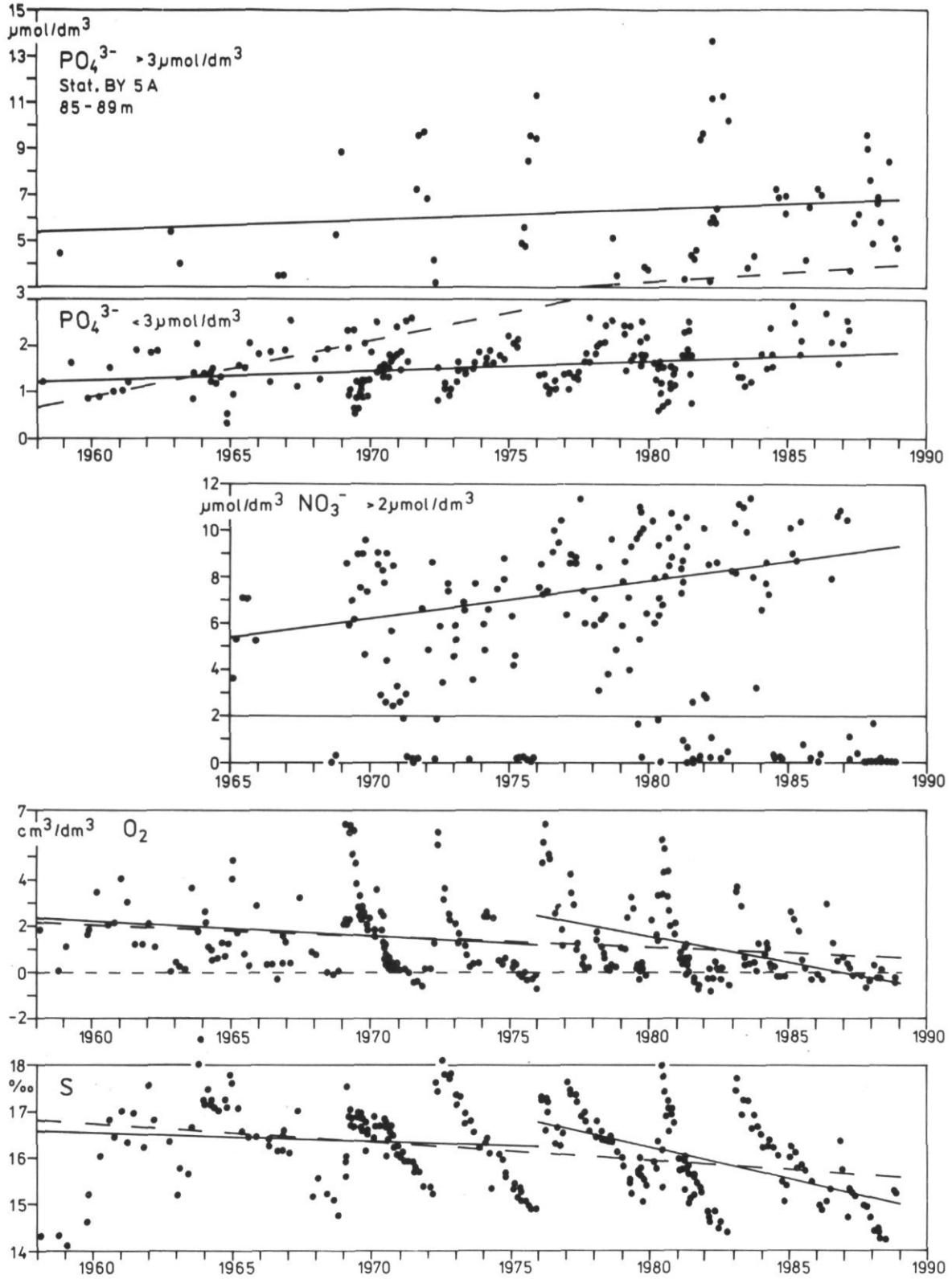

Figure 5. Variations and long-term trends of hydrochemical parameters in the near-bottom water layer of the Bornholm Deep (Station 5A, 85–89 m depth). Comments, see Figure 4.

phication is the deterioration of the oxygen conditions in Baltic deep waters indicated not only by the decreasing oxygen concentrations and accelerated oxygen consumption but also by increasing concentrations of hydrogen sulphide. This process is the result of the greater amounts of biogenic material produced in the euphotic layer mainly during the spring bloom of the phytoplankton.

The increasing phytoplankton biomass in polluted coastal areas in the Baltic Sea also reduces the penetration depth of light, causing the decrease of benthic macrophytes in shallow waters. In this way, spawning substrate is lost and fish larvae and young fish lose their nursery areas.

Nutrient discharge in coastal areas also favours exceptional algal blooms. An example was the toxic *Chrysochromulina polylepis* bloom observed in the Kattegat and North Sea in May 1988 (Dahl *et al.*, 1990).

Semi-enclosed seas located in the humid and arid climatic zones react in a different manner to land-based and airborne pollution owing to differences in the water balance. The water exchange at the entrances to the Baltic Sea, the Danish Straits, is characterized, on average, by the outflow of water with the lower salinity in the surface layer, whilst deep water with the higher salinity enters the Baltic Sea. The situation is the inverse at the entrance to the Mediterranean Sea; this means inflow in the surface layer and outflow in the depth of the Strait of Gibraltar (Fig. 6).

In comparison with the Baltic Sea and the adjacent Atlantic Ocean, the open Mediterranean is a highly oligotrophic sea (Coste *et al.*, 1988), although some coastal areas, mainly in the north and in the east are heavily influenced by sewage (McGill, 1965; Österberg and Keckes, 1977). The oxygen saturation is relatively high not only in the upper layers but also near the sea floor.

Nutrients in municipal and industrial discharges, although of great local significance in coastal areas, should have little effect on the trophic level in the Mediterranean Sea as a whole. After passing the biochemical cycle, the nutrients never return to the surface layer except by locally and seasonally restricted upwellings. They leave the Mediterranean through the Strait of Gibraltar after a certain residence time or are lost by geochemical sinks like deposition in the sediments. The inflow in the surface layer of the Strait of Gibraltar is relatively poor in nutrients and does not contribute to the eutrophication. This concept is also valid when the total amounts of phosphorus and nitrogen compounds are considered (Coste *et al.*, 1988).

The exchange at the entrances to the Baltic Sea is quite different. The nutrient concentrations leaving this sea through the Danish Straits in the surface layer are low except during the winter, whereas considerable amounts of nutrients are transported by the deep water current into the Baltic. These amounts are increasing due to the pollution of the North Sea. Figure 6 shows the differences in the phosphate exchange between semi-enclosed basins and the adjacent seas in the humid (Baltic) and the arid (Mediterranean) climatic zones.

Overall budgets show that only 10% of the nutrients originating from anthropogenic sources leave the Baltic through the Straits (Wulff and Stigebrandt, 1989). The rest remains causing eutrophication or is lost by internal biogeochemical sinks like deposition in the sediments or denitrification at low redox potential.

Because of differences in the water exchange, the pollution resulting from anthropogenic sources causes a nutrient accumulation which is much stronger growing in the semi-enclosed seas of the humid climatic zone as in those of the arid zone. In comparison to the Mediterranean, the Baltic is therefore much more sensitive against the increasing nutrient load.

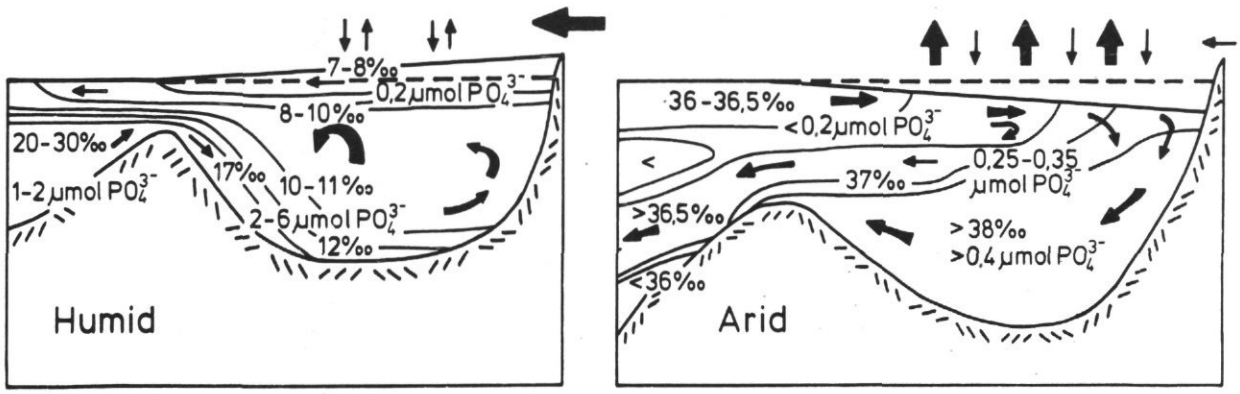

Figure 6. Water exchange between semi-enclosed basins and adjacent seas in different climatic zones (schematized). Humid zone: Great river run-off; evaporation equals precipitation (Baltic Sea). Arid zone: High evaporation, little precipitation and river run-off (Mediterranean Sea).

This conclusion is limited to the open sea areas. In the coastal areas of the Mediterranean and the Baltic, sewage is likewise a major problem. Therefore the countries bordering these seas must undertake great efforts in sewage treatment in order to protect the coastal waters against irreversible changes and to avoid polluted beaches.

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